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**Abstract**

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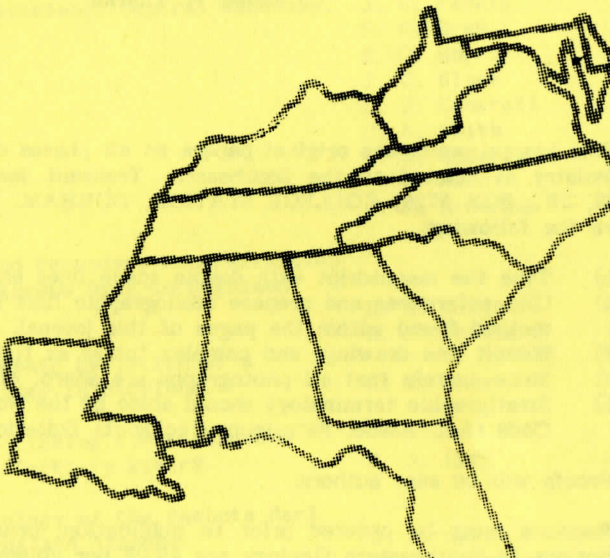
# SOUTHEASTERN GEOLOGY



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**SOUTHEASTERN GEOLOGY**

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# PALEOECOLOGY AND PALEOENVIRONMENTS OF THE BRYOZOAN-RICH SULPHER WELL MEMBER, LEXINGTON LIMESTONE (MIDDLE ORDOVICIAN), CENTRAL KENTUCKY

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## ABSTRACT

The Sulphur Well Member of the Lexington Limestone is unique in that bryozoan limestones compose over 50 percent of the member. The member represents a bryozoan biostrome that developed in a shallow, open-marine transition zone between a structrually controlled shoal and deeper, open-marine environments. Proximity to the shoal and to a sub-Sulphur Well disconformity, as well as possible growth faulting related to structural activity on the Kentucky River Fault System, appear to have exerted control on sediment and faunal distribution. In particular, lithofacies, diversity, and zoarial morphology can be related to each other and to proximity to the shoal. A wavy bedded calcirudite lithofacies generally formed closer to the shoal and is characterized by lower diversity and stout, ramose zoaria. A lenticular calcarenite lithofacies, on the other hand, was deposited more distal to the shoal, exhibits higher diversity, and is dominated by massive encrusting and platy foliaceous zoaria.

Overall, diversity in the Sulphur Well, except for bryozoans, is low compared to other members of the Lexington Limestone. In part, this apparently was related to the frequency of storms which periodically inundated Sulphur Well environments with sheets of sand washed from the adjacent shoal. Also, the bryozoans appear to have benefited from storm-generated breakage, burial, and transportation in ways that enabled them to rapidly colonize new substrates and exclude most other fauna.

## INTRODUCTION

The inner Blue Grass region of central Kentucky is underlain by the fossiliferous, Lexington Limestone, most of which is thin- to medium-bedded, gray, bioclastic limestone and interbedded shale. The Sulphur Well Member of the Lexington Limestone, however, is unique, because its principal lithologic components are extremely abundant fossil bryozoans. By definition (Cressman, 1972), over 50 percent of the unit is composed of bryozoans. Megascopically, the unit is best described as a biostromal bryozoan limestone or byozoan calcirudite. The purpose of this study is to determine the origin of the Sulphur Well Member. Namely, what were the paleoenvironmental and paleoecological conditions that permitted so many bryozoans to live in such a local area?

## PROCEDURES

A distribution map of the Sulphur Well Member (Figure 1) was made by compiling outcrop patterns from 16 U. S. Geological Survey geologic quadrangle maps in central Kentucky. The quadrangle maps were used to select 16 exposures (Figure 1) which were discribed, measured, and photographed using standard field procedures. Stratigraphic relationships between the

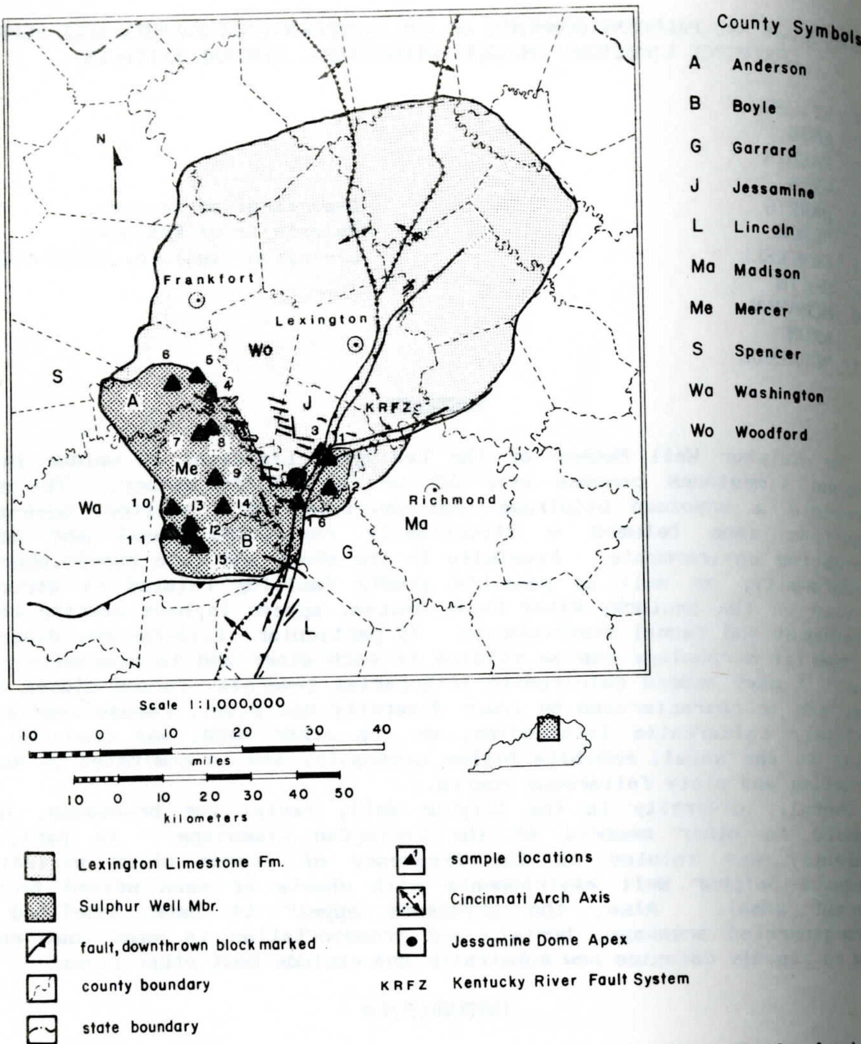


**Figure 1. Location map showing approximate distributions of the Lexington Limestone and Sulphur Well Member, the location of studied exposures, and the location of major structures in the area. The invagination in the Sulphur Well distribution pattern near the apex of Jessamaine Dome is the position of the Tanglewood shoal.**

Sulphur Well and adjacent units were noted, as were all biogenic and sedimentary sections. Representative lithologies and sedimentary structures were collected for sectioning and photographs. Data collected in the field and from geologic maps were used to construct structural cross sections, an isopach map (Figure 2) and a paleogeologic map of the sub- Sulphur Well surface (Figure 3).

Fossils were collected and identified, and where possible, their taphonomic condition, external morphology, and substrate occurrence were noted for possible paleoecological implications. Because of difficulty in identifying bryozoans, zoarial morphology was noted and specimens were identified to the ordinal level; species lists were obtained from the literature.

The literature was examined for information on stratigraphic history, as

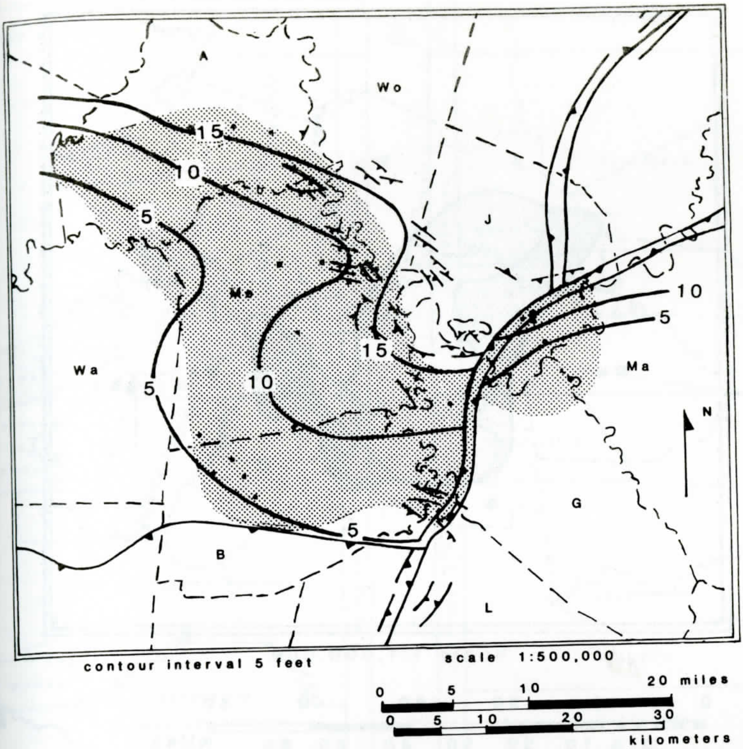


**Figure 2. Isopach map of the Sulphur Well Member superimposed on the surface distribution pattern. Outcrop pattern on the northern margin of unit distribution exhibits an invagination that coincides with the Jessmaine Dome (Figure 1) and parts of the Kentucky River Fault System suggesting structural control. County abbreviations same as those on Figure 1.**

well as for information on the structural, tectonic, paleoclimatic, and paleogeographic settings. This information was integrated with data from the field to produce the paleoenvironmental and paleoecologic interpretations presented herein.

## STRATIGRAPHY

### Previous Work

The Lexington Limestone is a Middle-Upper Ordovician unit originally established by Campbell (1898) and subsequently expanded upward by Miller (1917, 1919) to the base of the Cynthiana Formation (Figure 4B). Miller (1917, 1919) described five members based on faunal content and lithology and interpreted them to be tabular units with widespread distribution. McFarlan (1943) and McFarlan and White (1948) expanded and subdivided the Lexington Limestone further and demonstrated complex facies relationships between some of the members; they still, however, maintained an essentially tabular interpretation for most of the members, an interpretation reinforced by the incorrect usage of biostratigraphic zones as lithologic units (Gutstadt, 1958). This usage continued until the 1960's and 1970's when the joint U. S. Geological Survey-Kentucky Geological Survey mapping program showed that few, if any, of the Lexington members exhibited tabular geometry. Instead, the new mapping showed these members to be part of a complexly intertonguing facies mosaic related to periodic uplift on structures in the Lexington area

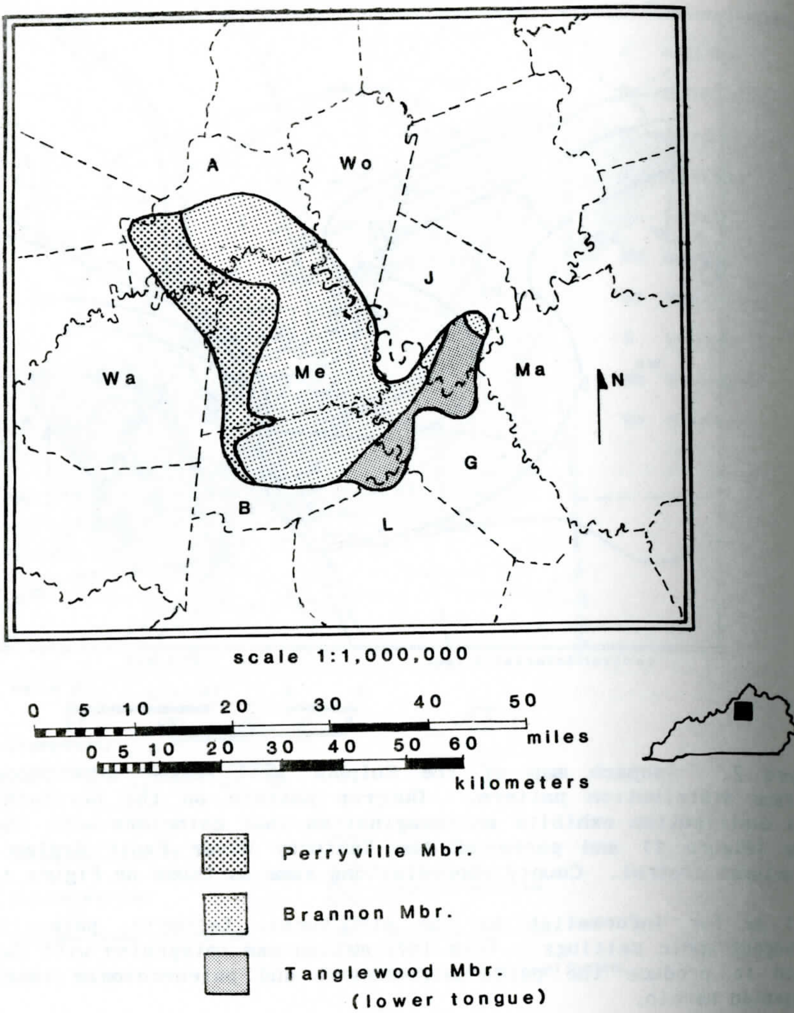


**Figure 3. Paleogeologic map of the disconformity surface below the Sulphur well. The Tanglewood shown is the lower tongue of the Tanglewood Member.**

(Black and others, 1965; Cressman, 1973). Use of unit names with biostratigraphic connotations was discontinued, much of the Cynthiana Formation was included with the Lexington (Figure 4), and new lithostratigraphic units were designated (Weir and Cressman, 1978). One of the older units whose designation was changed is the Sulphur Well Member.

The bryozoan-rich rocks that comprise the Sulphur Well Member were originally known as "pre-Greendale" because they were present locally below the Greendale Member of the Cynthiana Formation (McFarlan, 1938), or as the *Crepipora* zone because they contain the characteristic bryozoan species *Crepipora spatiosa* (McFarlan, 1938, 1943; McFarlan and White, 1948; Nosow and McFarlan, 1960). In 1943, McFarlan formally designated the Sulphur Well as the basal member of the Cynthiana Formation in southern and southwestern parts of the inner Blue Grass area; it was subsequently assigned to the Lexington Limestone by Cressman (1968).

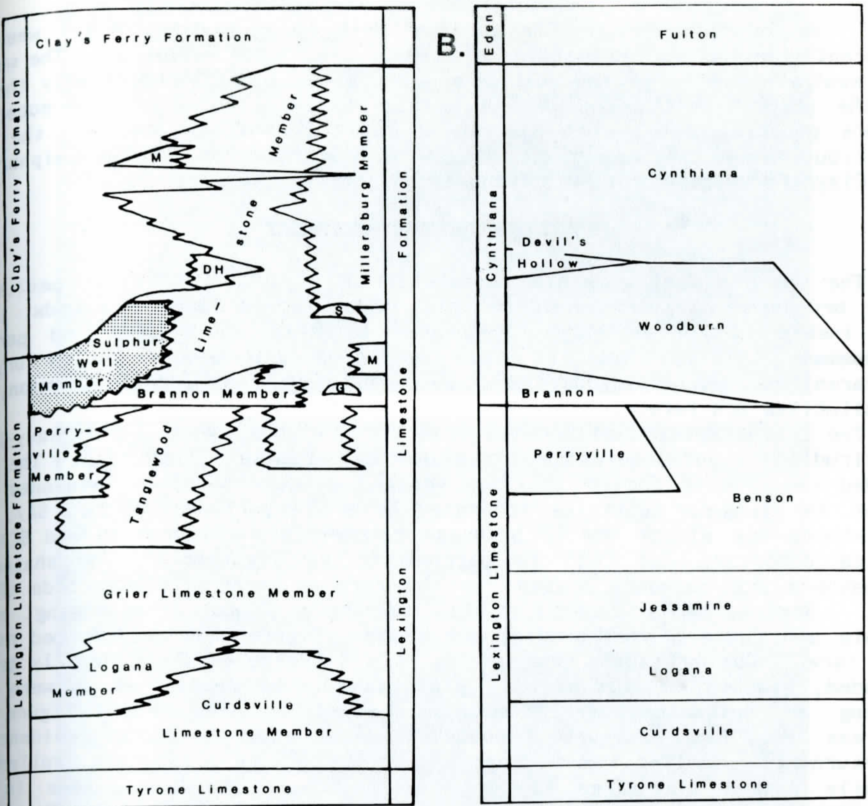


**Figure 4. Presently accepted stratigraphic relationships and nomenclature for the Lexington Limestone (after Cressman, 1973) (A), compared with those of Nosow and McFarlan (1960) (B). In the more recent interpretation three tongues of the Tanglewood are present; a lower tongue below the Sulphur Well, a middle tongue equivalent to the Sulphur Well, and an upper tongue, above the Sulphur well. Abbreviations: M, Millersburg Mbr.; S, Strodes Creek Mbr.; G, Greendale Lentil; and DH, Devils Hollow Member.**

## Characteristics

The Sulphur Well is an upper Middle Ordovician unit (Caradocian; Shermanian) that conformably underlies the interbedded shales and limestones of the Clays Ferry Formation (Figure 4A). North of central Mercer County, this contact is sharp, but to the east and south of this area the contact is vertically and laterally gradational from bryozoan-rich limestone in the Sulphur Well to bryozoan-rich shale in the Clays Ferry (Cressman, 1973). The lower contact is sharp, planar, and largely disconformable with underlying Lexington members. The disconformity truncates progressively older members to the west, so that the Sulphur Well overlies the lower tongue of the Tanglewood Member in the east, the Brannon Member in the central area, and the Perryville Member in the west (Figures 3 and 4A).

The extent of the Sulphur Well is shown in Figures 1 and 2. Thickness of the unit varies from 16 ft. (5 m) in the extreme north to less than five feet (1.5 m)in the south. To the north and east, the unit abruptly disappears as it grades into coarse-grained calcarenties and calcirudites of the middle tongue of the Tanglewood Member (Figure 4A) or Tanglewood-like lithologies mapped with the Grier (Wolcott, 1970; Cressman, 1973); the lower and upper tongues of the Tanglewood Member (Figure 4A) are not associated with the Sulphur Well. The Sulphur Well thins southward and westward, apparently by lateral gradation into the Clays Ferry Formation. The western and southern limits of the unit shown in Figures 1 and 2 define only the edge of the outcrop belt. The Sulphur Well probably extends into the subsurface before pinching out within the Clays Ferry Formation; however, the exact distribution in the subsurface cannot be discerned because the Sulphur Well and Clays Ferry give similar responses on geophysical logs.

## LITHOLOGY AND SEDIMENTOLOGY

The sulphur Well consists dominantly of light gray (N7) to medium-gray (N5) bryozoan-rich limestones in thin, irregular to lenticular beds, two to six inches (3-15 cm) thick, separated by thin shale beds and partings (Cressman, 1973). Most of the limestones are poorly sorted bryozoan calcarenites and calcirudities; calcisiltites are locally common, and calcilutites are rare.

Two lithofacies are common in the Sulphur Well: a wavy-bedded calcirudite lithofacies and a lenticular calcarenite lithofacies. The wavy-bedded calcirudite facies is characterized by more massive limestones which occur in elongate lenticles separated by shale partings (Figure 5A). The limestones are almost wholly bryozoan calcarenites (Figures 5B and 5C), and shales comprise less than 10 percent of any exposure. The shales and limestones are commonly present in the form of wavy or flaser beds (Figure 5A). Scours, small channel fills, low-angle planar crossbedding, rip-up clasts and crude graded bedding are common (Figure 5C); well-formed ripples are rare. The bryozoans present in this lithofacies are generally broken, abraded, stacked or imbricated. Platy foliaceous zoaria, which were free-living or unttached, are commonly stacked or imbricated (Figure 5B), whereas the few massive, "knobby" zoaria always exhibit evidence of overturning. Smaller ramose zoaria apparently were fragmented, rolled, and tightly packed like logs (Figure 6A). Exposed bedding planes in this lithofacies reveal nearly complete, crushed "bushy" colonies (Figure 6B) which seemingly experienced little transportation. Apparently, such colonies were buried in place and broken during later compaction. Locally, undeformed, bifoliate platy zoaria (Figure 6C) are abundant. This lithofacies is most common near the northern and eastern boundaries of the member where it intertongues with the Tanglewood.

The lenticular calcarenite lithofacies is characterized by thinner-bedded, finer-grained limestones in small lense-like bodies (Figure 7A). The limestones are dominantely calcarenites and calcisiltites, and interbedded shales are thicker, generally comprising 30 to 40 percent of an exposure. The small, lens-like limestones (Figures 7A and 7B) are usually calcarenites, and together with adjacent shales give rise to lenticular bedding, commonly with connected lenses (Figure 7B; see Reineck and Singh, 1980, p. 115). The presence of micro-cross-laminae in these lenses indicates that most originated as ripples on the distal edges of sand sheets. The shapes of the ripples have been subsequently modified by encrusting bryozoans, bioturbation and compaction. Small calcarenite bodies with a channel-fill geometry also are present (Figure 7B). The nodular limestones are calcisiltites and rare calcilutites associated with thicker shale beds. Massive, "knobby" and platy foliaceous zoaria are found closely associated with the calcarenite lenses; some of the massive zoaria have been overturned. Fragmented ramose, frondose and platy zoaria also are found tightly packed in intervening shales. This lithofacies is more common in southern and western parts of the outcrop belt, where it is associated with the more shaly Clays Ferry Formation.

Thin-section petrography of Sulphur Well limestones indicates that most are poorly sorted crinoid-bryozoan-fragment grainstones and packstones. Grainstones clearly dominate. The packstones exhibit an argillaceous micrite matrix which is locally replaced by pseudospar, dolomite , and phosphate. Phosphate is common in all the sections, generally replacing echinoderm

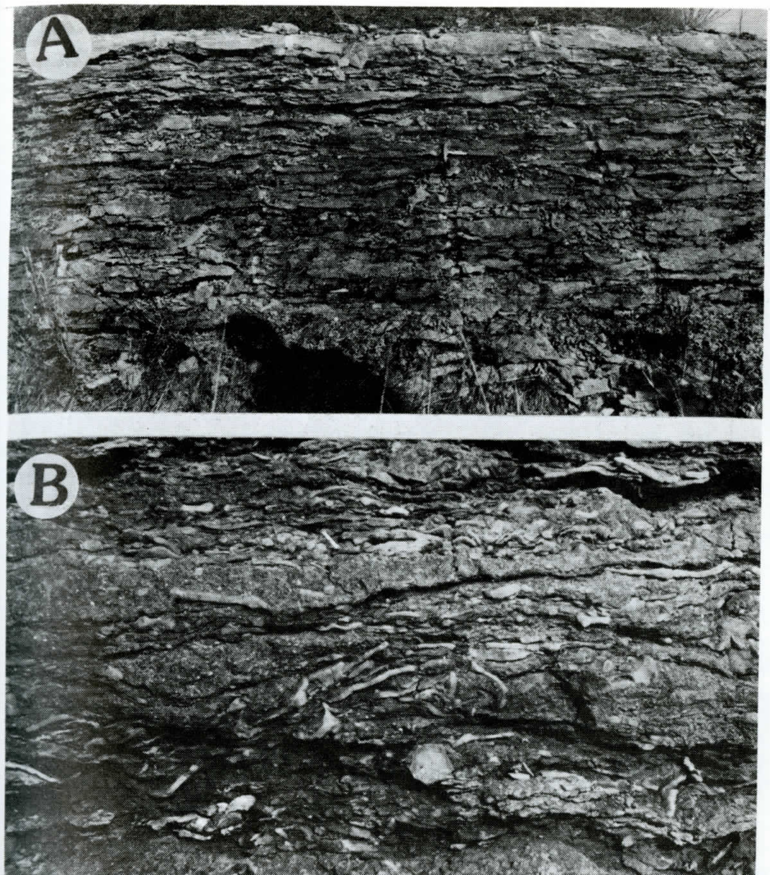

Figure 5. A. Wavy-bedded calcirudite lithofacies from location 5 (Figure 1). B. Stacked and imbricated fragments of platy zoaria in wavy-bedded calcarenites and calcirudites from location 8 (Figure 1). Bryozoan fragments in center are approximately (2.5 cm) in length. C. Poorly graded channel-fill sequence of bryozoan calcirudite and calcarentie from location 8 (Figure 1).

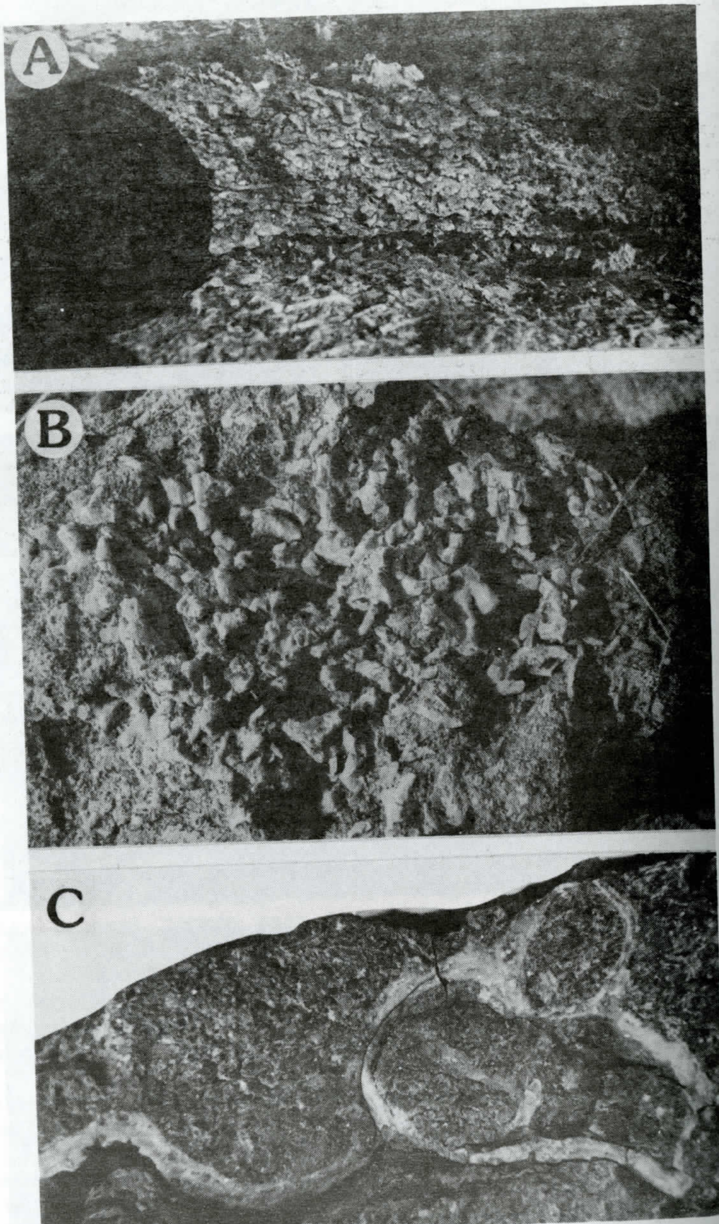


Figure 6. A. Fragmented ramose zoaria in a bryozoan calcirudite stacked like "logs". A largely in-place broken colony is present on the bedding plane below; from location 5 (Figure 1). B. Elliptically shaped ramose colony on the top of a bedding plane from location 4. The colony at one time probably exhibited a "bushy", rounded form which was the result of and facilitated a tumbleweed-like transportation along the bottom during storms. Long axis of colony is approximately 8 in. (20 cm.) long. C. Platy, foliaceous, bifoliate zoarium from location 14 (Figure 1). Note the undulatory nature of the zoarium and the infilling of all "cavities" with sediment which prevented crushing. Length of specimen is 4.5 in. (11.4 cm.).

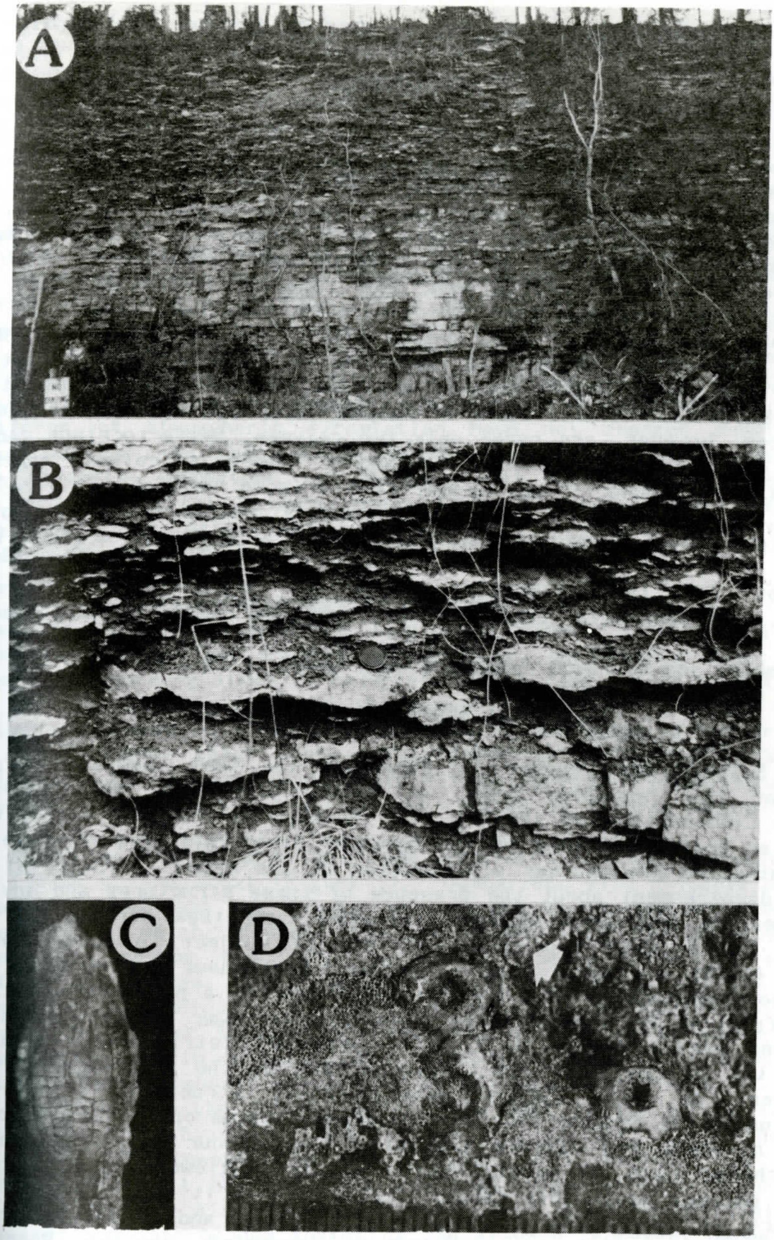

Figure 7. A. Exposure of the lenticular calcarenite lithofacies in the upper half of the exposure at location 1 (Figure 1). The nodular base of the member intertongues with more massive Tanglewood lithologies at the middle right. B. Close-up view of the lenticular calcarenite lithofacies from location 2 (Figure 1). Note the lenticular bedding of calcarenite lenses in shale. Platy zoaria rest on the calcarenite lenses and together with smaller ramose zoaria form a dense biostromal network in intervening shale. Calcarenite body in the lower right is part of a small channel. C. *Dystactocrinus constrictus* from location 6 (Figure 1); crinoid is 1.1 in (2.8 cm) high (D) unidentified crinoid holdfasts and *Zygospira* (arrow) on a massive, "knobby" zoarium from location 1; scale in centimeters.

ossicles, bryozoan fragments, and the micrite matrix.

## PALEONTOLOGY

Fossils are very abundant in the Sulphur Well, but overall diversity is low, except for the bryozoans which make up most of the fauna. The trepostomes, *Homotrypa*, *Pernopora*, *Prasopora*, *Heterotrypa*, *Dekayia*, *Eridotrypa*, *Batostoma*, *Cyphotrypa*, *Parvhallopora*, *Stigmatella* and *Hemiphragma*, the cystoporates, *Constellaria*, *Crepipora*, *Acanthoceramoporella* and *Ceramoporella*, as well as the cryptostomes *Escharopora*, *Graptodictya*, *Trigonodictya*, and *Pachydictya* have been reported (McFarlan, 1938; McFarlan and White, 1948; Cressman, 1968; Wolcott, 1969, Karklins, 1984).

Crinoids are the next most common fossil type in the Sulphur Well, but based on the number of species indicated by calyces and holdfasts, their diversity was low. Disarticulated ossicles are everywhere abundant and are one of the primary constituents of Sulphur Well limestones. Intact columns and calyces, however, are rare; only two calyces, *Dystactocrinus constrictus* (Figure 7C) and *Columbicrinus* (?), and a number of columns encrusted by bryozoans were found. An unidentified crinoid holdfast (Figure 7D) was commonly found encrusting platy and massive zoaria. One example of a *Glyptocrinus* stem apparently coiled around a bryozoan zoarium was also found.

Except for the ubiquitous *Zygospira*, brachiopods are relatively uncommon. *Herbertella frankfortensis* and *Rafinesquina* were found only locally.

Trilobite and ostracod fragments were noted in thin section.

## STRUCTURAL AND TECTONIC FRAMEWORK

The Sulphur Well Member is found on the west, south, and southeast flanks of the Jessamine Dome, an upwarp along the northeast-southwest trending Cincinnati Arch (Figure 1). Erosion on the arch and on the dome, as well as uplift on parts of the Kentucky River Fault System, have in large part controlled present-day distribution of the member. There is, however, general disagreement about the presence of these structures and any possible synsedimentary influence during Lexington time (see Borella and Osborne, 1978). In the most thorough study on the subject to date, Borella and Osborne (1978) indicated that an active, continuous arch was not present. They also demonstrated that the Jessamine Dome or a precursor, controlled by growth faulting on the Kentucky River System, was present and influenced deposition during Lexington time. Based on distributional relationships between the Sulphur Well and Tanglewood and on the location of present-day structures, we also suggest synsedimentary structural activity during deposition of the Sulphur Well and the middle tongue of the Tanglewood.

The northern and eastern limits of the Sulphur Well are defined by a facies change into the middle tongue of the Tanglewood Member (Figure 4A). What is unusual about this facies change is that it approximately coincides with part of the Kentucky River Fault System and a smaller northwest-southeast fault system in southwestern Jessamine, southern Woodford, and western Mercer counties (Figures 1 and 2). In both fault systems, the downthrown side is dominantely to the south. The coincidence of the facies change and the fault systems suggests synsedimentary structural control by growth faults. The high-energy Tanglewood shoal apparently developed on the uplifed northern area, whereas shallow open-marine Sulphur Well environments formed in probable downthrown areas to the south. Moreover, Sulphur Well isopachs are sub-parallel to the trends of the faults (Figure 2), and the fault-bound Tanglewood shoal area roughly coincides with the apex of the Jessamine Dome. The apex is reflected in the prominent south-trending reentreant in the Sulphur Well outcrop pattern in northern Garrard and southwestern Jessamine counties (Figures 1 and 2).

The Sulphur Well outcrop pattern to the south and southwest is clearly controlled by surface faults (Figure 2), but the absence of facies changes here indicates that the faults probably exerted no control over the deposition of the unit. The Sulphur Well is apparently present south and east of the faults but downdropped below the level of erosion.

## PALEOGEOGRAPHIC AND PALEOCLIMATIC SETTING

During the late Middle Ordovician, most of east-central United States was the site of shallow-water carbonate deposition (Cressman, 1973). Large areas of even shallower, shoal environments related to synsedimentary uplift on basement structures apparently formed contemporaneously on or around the Jassamine, Nashville and Ozark domes (Ervin and McGinnis, 1975; Borella and Osborne, 1978). Based on the paleogeographic reconstructions of Ziegler and others (1979) and Scotese and others (1979), central Kentucky at this time was located approximately 15 degrees south of the paleoequator. Assuming that patterns of atmospheric circulation have not changed substantially throughout much of the Phanerozoic (Drewry and others, 1974), we surmise that the Lexington Limestone and Sulphur Well Member were deposited in tropical conditions within a warm, evaporative, trade-wind belt. This means that the prevailing paleowinds approached the Lexington area from the present-day east or northeast.

## PALEOENVIRONMENTAL INTERPRETATIONS

Unlike the laterally contiguous Tanglewood Member, or the overlying and underlying Clays Ferry and Brannon respectively, the Sulphur Well contains both fine- and coarse-grained lithologies. This combination of lithologies and the stratigraphic position of the member (Figure 4A) suggest that the Sulphur Well developed in a transitional area in the sense of Multer (1971): between a high-energy shoal system (Tanglewood on the Jessamine Dome) and quiet, deeper, open-marine environments (Clays Ferry). Transitional areas typically are subject to tidal action and limited in lateral extent (Multer, 1971).

The extent of the Sulphur Well was similarly limited to western and southern flanks of the shoal (Figure 1). Comparison with probable paleowind directions indicates that the Sulphur Well formed on the leeward side of the Tanglewood shoal, which afforded the environment and its bryozoan fauna protection from storms and large wind-driven waves from the east and northeast. Nonetheless, the gradient between the shoal and the leeward transition area periodically allowed transportation of many shoal-derived sand sheets into deeper Sulphur Well environments (Figure 8) during storms. Consequently, the origin of the Sulphur Well Member is closely tied to that of genetically related shoals represented by the Tanglewood Member.

More specifically, the Sulphur Well is related to the middle tongue of the Tanglewood with which its intertongues (Figure 4A). Both the middle tongue of the Tanglewood and the Sulphur Well overlie the Brannon disconformably where it has not been completely removed by erosion. Basal parts of both members also commonly contain reworked Brannon clasts. The calcisilitites and interbedded shales of the Brannon represent deposition in a quiet, deeper-water environment. However, the abrupt appearance of the middle tongue of the Tanglewood and Sulphur Well, as well as the erosion of the Brannon below them, probably reflect uplift in the Tanglewood shoal area. Although the middle tongue of the Tanglewood is largely restricted to an uplifted area defined by the Jessamine Dome and associated fault systems, submarine erosion associated with the uplift apparently extended seaward to the south and west forming the disconformity below the Sulphur Well.

Following this uplift and erosion, deposition resumed in environments that apparently were shallower than those represented by the Brannon. Deposition was generally controlled by proximity to the shoal. Closer to the shoal, the coarser-grained, wavy-bedded calcirudite lithofacies dominates

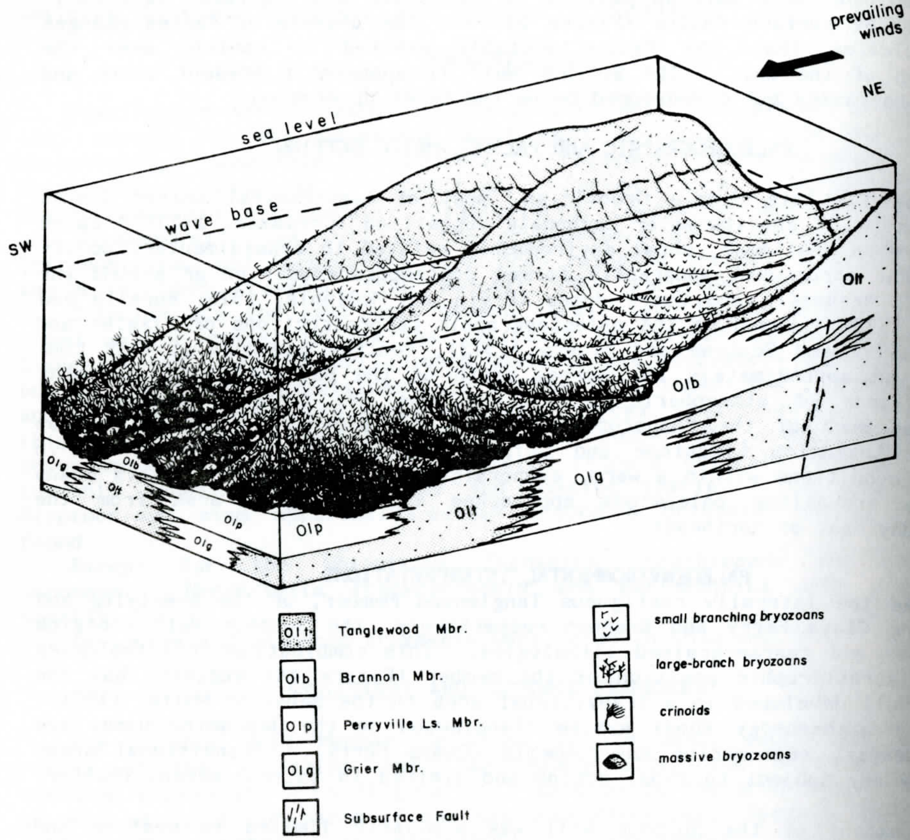


**Figure 8. Block diagram showing environmental reconstrution of Sulphur Well and its stratigraphic relations with other members. Sulphur Well occurs leeward of Tanglewood shoal (stippling). Light area below wave base is the wavy-bedded calcirudite lithofacies; darkened area with massive bryozoans is the lenticular calcarenite lithofacies. No scale intended.**

(Figures 5A, B, and C). Even though this part of the Sulphur Well probably was below wave base, during storms bryozoan colonies apparently were broken and transported basinward along with spillover lobes of coarse, bioclastic sand from the Tanglewood shoal (Figure 8). The crudely graded, bioclast-dominated calcirudites (Figure 5C) of this lithofacies are characteristic of storm-generated tempestites (Aigner, 1982). The elongate lenticles (Figure 5A) are merely sectional views across various parts of stacked sand sheets that were transported seaward from the Tanglewood shoals. The wavy nature of the beds is a product of scouring at the bases and modified rippling at the tops; both types of features are emphasized by shale partings. Rippling may be formed by waning storm-generated currents and waves (Aigner, 1982) or by long-shore reworking of shallower spillover sand sheets (Imbrie and Buchanan, 1965; Ball 1967).

The shale partings in this lithofacies (Figure 5A) are most likely the product of normal pelagic sedimentation; much of the mud may have been put into suspension during storms. Erosion during storms and scouring during emplacement of sands probably has drastically reduced the thickness of the shales.

More distal to the shoal, the lenticular calcarenite lithofacies dominates. In this area, only the finer-grained, thinner, distal edges of the storm-generated sand sheets or spillover lobes are found. Ripples are present on the sand sheets, but are commonly incomplete (Figure 7B). This was probably caused by meager sand supply (Reineck and Singh, 1980), which also explains the abundance of lenticular bedding in this lithofacies (Figures 7A and 7B), the presence of bioturbation, and the abundance of nearly in-place platy, ramose, and massive bryozoan zoaria. Bioturbation and growth of platy bryozoan zoaria, however, have substantially altered the surface form of many rippled sand sheets.

In the southwestern part of the outcrop area, the Sulphur Well is interrrupted by a tongue of unfossiliferous Clays Ferry lithology (Cressman and Karklins, 1970, Figure 8; Cressman, 1974). The calcisiltites, calcilutites, and shales of this tongue reflect a period of even deeper, more distal conditions and show the proximity of transitional Sulphur Well areas to deeper, Clays Ferry environments.

## PALEOECOLOGICAL INTERPRETATIONS

### Regional Controls

The great abundance of bryozoans in the Sulphur Well probably is the most important charactertistic of the member. The bryozoans trapped and stabilized sediment and were major sediment contributors; yet their localized abundance in the member is poorly understood. We suggest that this abundance is related to the paleogeographic and paleoclimatic setting of the Tanglewood shoal around which the Sulphur Well developed. The relatively shallow (10-70 m), clear, slightly to moderately agitated waters that characterize such environments in a tropical setting support the richest bryozoan assemblages (Cuffey, 1970). In addition, the presence of phosphate indicates incursion of at least periodic upwellings (Cressman, 1973). More important, however, was the transitional nature of Sulphur Well environments on the leeward side of the Tanglewood shoal where they were protected, and at the same time, received waters rich in oxygen and nutrients transported beyond the shoal by tidal currents. Similar leeward environments support prolific bryozoan communities today on the Bahama platform (Newell and others, 1959; Hoffmeister and others, 1967) and apparently did so in the geologic past as well (Hoffmeister and others, 1967; Multer, 1971; Cuffey, 1977; McKinney, 1979; McKinney and Gault, 1980).

### Effect of Storms

Environmental hazards, however, also exist in such transitional environments, and of these, storms are most significant. During storms, sheets of sand are washed off the shoals and transported into adjacent environments, burying many invertebrate communities. This, we believe, was the origin of most of the rippled sand sheets (Figures 7B and 8) in the Sulphur Well. Not all the bryozoan colonies, however, were buried by sand sheets. In the lenticular calcarenite facies, many platy foliaceous zoaria were found interlaminated with shale, suggesting that they were buried by muds which may have been resuspended by storms. Moreover, during storms, wave base may been lowered to bottom level, resulting in breakage and transportation of skeletal elements. The stacked and imbricated bryozoan fragments (Figures 5B and 6A) and overturned zoaria probably originated in this way.

The storm also apparently had some beneficial effects. Old, crowded, and successionally stable communities were periodically buried by the sand sheets, and these sheets, in turn, provided new substrates for recolonization and reinitiation of community succession. Such disturbances, in fact, may have maintained increased species diversity by opening space for opportunistic, early, successional forms and preventing monopolization of space by competitively dominant, later, successional forms (Wilson, 1985).

Most of the new substrates probably were recolonized rapidly, because they were littered with fragmented zoaria capable of regeneration after transportation. Regeneration of colonies from fragments is common in corals (Highsmith and others, 1980; Tunnicliffe, 1981; Highsmith, 1982), as well as in bryozoans (Blake, 1976; Winston, 1981, 1983; Cheetham and others, 1981; McKinney, 1983) and apparently was an important reproductive mode. The presence of frondose branches emanating from the upper surfaces of fragmented platy zoaria lying on former muddy Sulphur Well substrates suggests that this mode of reproduction was common in Sulphur Well environments. Moreover, the apparent rapidity with which the substrates were recolonized indicates the importance of colony fragmentation in the Sulphur Well, only a few tens of millions of years after the beginning of the fossil record of bryozoans and much earlier than previously documented. Hence, the periodic storms seem to have been important agents of colony reproduction and dispersal in the Sulphur Well. Judging from the nature of colonies on superimposed sand sheets, the time between recurrent storm events at any one locality was long enough to permit complete recolonization of the substrates.

Throughout the Sulphur Well, fauna and lithology appear to be closely associated. Overall, the most diverse assemblages are associated with the finest-grained lithologies, which in turn are thickest where the sub-Sulphur Well disconformity is best developed. Apparently the deeper, more protected Sulphur Well environments were associated with erosional lows on the bottom (Figure 8).

## Zoarial Morphology And Significance

Zoarial morphology also was closely related to lithology. The knobby, massive zoaria are restricted to the calcisiltite and shale lithologies in the lenticular calcarenite lithofacies. We believe that the large, massive zoaria were restricted to the finer-grained lithologies because zoarial growth was able to keep pace better with the slower rate of sedimentation characterizing these areas. As a result, some part of the zoarium was always above the muddy substrate, whereas on coarser substrates, such zoaria were more likely to be buried or overturned.

Overturning by itself, however, was not necessarily detrimental to a colony, and probably was instrumental in the development of the massive habit. Many of the massive, knobby zoaria appear to have formed by successive encrustation of thickly branched zoaria. Through overturning, many of the branches were broken forming a rounded, cobble-like nucleus that was easily encrusted. In fact, these massive zoaria are the only examples of encrustation in the entire Sulphur Well, probably because they provided the only hard substrates available. The knobby nature of the zoaria is the result of encrustation of the broken branches. Those ends not rapidly encrusted were heavily fouled, largely by borers (Figure 9A). Overturning and breakage not only provided the rounded nucleus for encrustion, but also periodically killed surface encrusters and foulers and opened space for renewed community succession (Wilson, 1985). Through this accretion by encrusters, massive zoaria continued their growth until completely buried. The largest zoarium we found had a long demension of 21 cm and a height of 15 cm. Crinoid holdfasts and bores (Figures 7D and 9A) are found on all surfaces of these zoaria supporting the overturning hypothesis. In many ways, the massive, knobby zoaria are similar to the "ectoproctaliths" or "rolling stones" of Rider and Enrico (1979) and Dade and Cuffey (1984) and to the cobble-dwelling hardground fauna of Wilson (1985); they have the textural attributes of cruststones and bindstones (Cuffy, 1985).

Platy foliaceous zoaria were found in place within all lithologies except in the coarsest calcirudites (Figure 10), where they had obviously been reworked (Figure 5B). The generally flat, platy nature of such zoaria allowed these bryozoans to colonize unstable, coarse and soft substrates. Many of the zoaria also exhibit a wavy or undulatory, potato-chip-like

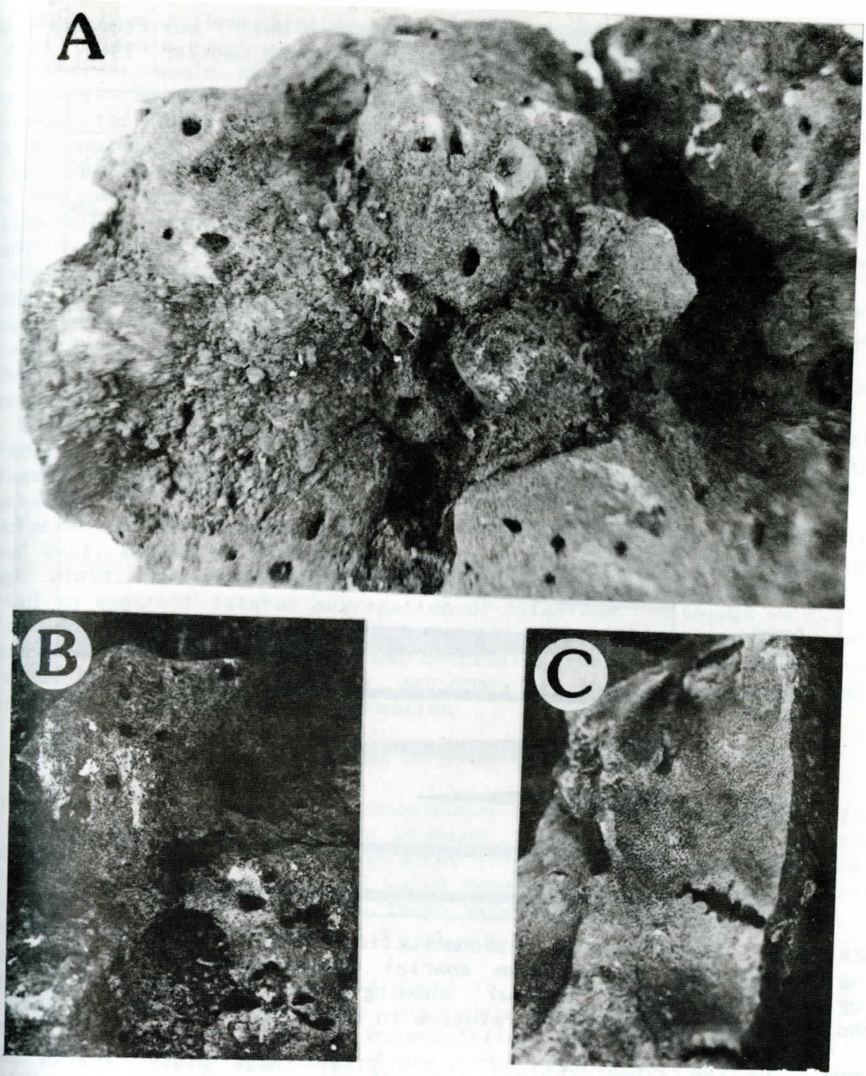

**Figure 9. A. Part of a massive, "knobby" zoarium showing surface fouling by borers and encrusters. The larger bores in the center, upper right and upper left occur in the centers of broken branches; crinoid holdfast in upper center. Part of zoarium shown is 4.7 in. (12 cm.) in long dimension. B. *Trypanites* - like borings from location 1 (Figure 1). C. "Zigzag", borings from location 8.**

morphology (Figure 6C). The presence of so many concave-up and convex-up surfaces ensured that no matter how the zoaria landed on a substrate after transportation, virtually some part would be exposed and continue to grow (F. K. McKinney, personal communication, 1985). Moreover, filling of all the cavities and undulations with sediment would have prevented crushing of the zoaria (Figure 6C) and enabled their use as holdfasts for parts of colonies still above the substrate. Many of the rippled sand sheets in the lenticular calcarenite facies are literally veneered with these platy zoaria, although they also are common in the intervening shales. Platy zoaria probably were the most susceptible to burial, either by migrating sands or settling mud, and as expected we found that they were more abundant in lithologies representing deeper, more distal Sulphur Well environments where sedimentation was slower (Figure 10) (see Lagaaij and Gautier, 1965,

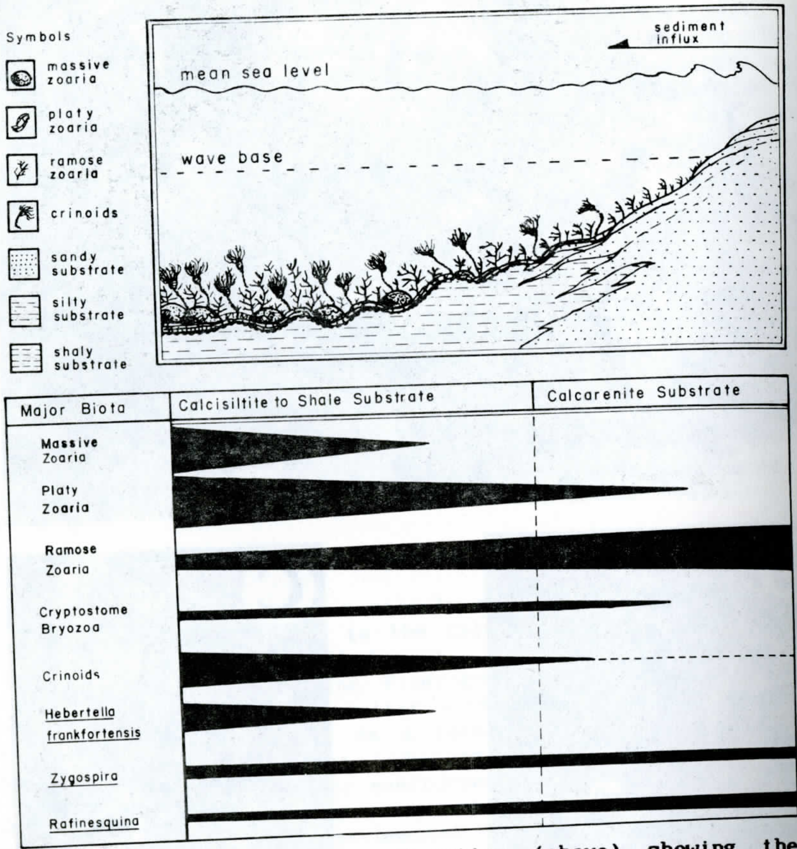


**Figure 10. Environmental reconstruction (above) showing the relative ecologic positions of bryozoan zoarial morphologies and a corresponding ecologic range chart (below) showing the distribution of zoarial morphologies and other fossils relative to substrate.**

p. 50). In the finest-grained lithologies, these platy, zoaria commonly form a dense, biostromal framework, which has the textural attributes of a bafflestone or lettucestone (Cuffey, 1985).

Rigid, ramose zoaria (Figure 6B) were found in all lithologies (Figure 10) as well, but they were most abundant and robust in the wavy-bedded calcirudite facies. Preference for these coarse substrates probably reflects a need for firm substrates and the ability to accommodate partial burial by mobile sands because of elevation above the bottom. Broken fragments are present everywhere; apparently complete ramose colonies, however, are found only on bedding planes (Figure 6B). Although crushed, these colonies apparently experienced little major breakage before burial.

Most of these ramose colonies exhibit a roughly elliptical or circular shape on the bedding planes (Figure 9B), which suggests a "bushy" spherical or ellipsoidal colony geometry during life. This geometry and the fact that holdfasts are never found suggest that the colonies were free and capable of rolling movement similar to wind-blown tumbleweeds; similar modes of transportation for bryozoans have been inferred by Cheetham and others (1981) and McKinney (1983). We believe that the above features reflect adaptation for life in high-energy environments. During storms, these bryozoans would

would have been rolled along the bottom like tumbleweeds. Although some of the more delicate growing tips of the colony probably were broken, the more rigid interior would have persisted to start new colony growth after the storm waned. No matter what part of the colony came to rest directly on the substrate, the irregular branches would have formed an ideal rhizoid-like means of anchorage and support. Moreover, it is likely that once colonies were so stabilized, they formed fixed obstructions in the path of normal sand-laden bottom currents resulting in small sand drifts, which further stabilized the colonies. Projecting branches near the base of the colonies probably helped to retard destabilizing sediment scour around the colony in much the same way as do spines on brachiopods (Alexander, 1984).

Smaller branching cryptostome zoaria are found associated with the massive and platy zoaria (Figure 10). They most likely formed small thickets nestled upon and between larger zoaria.

Overall, the distribution of zoarial forms in Sulphur Well environments reflects what was recently noted by Jackson and Hughes (1985), namely that more stationary, massive growth forms are generally more abundant in environments with low levels of disturbance, whereas mobile (through growth or fragmentation) forms are more abundant in environments with high levels of disturbance.

McFarlan (1938) reported a distinct vertical zonation of bryozoans in the Sulphur Well, beginning with ramose zoaria at the base and progressing upward through platy to massive zoaria at the top. In a few places, he also reported an apparent lateral segregation of *Batostoma*. Although we could not verify this zonation, the reported changes in zoarial types appear to reflect a trend toward increasing depth and decreasing energy. This same trend is indicated by lithologic changes accompanying the vertical and lateral transition into the Clays Ferry Formation.

## Implications Of Other Fauna

Crinoids also formed a very conspicuous part of the Sulphur Well fauna. They seem to have been especially abundant in the calcisiltites and shales of the lenticular calcarenite lithofacies (Figure 10), primarily because the massive and platy zoaria provided solid substrates for attaçhment. Crinoid holdfasts are extremely common on these zoaria (Figure 7D). However, only one or two types of holdfasts are found, indicating that only a few crinoid species were really abundant.

Articulate brachiopods are also found in the Sulphur Well Member but are much less conspicuous than the bryozoans and crinoids. *Herbertella frankfortensis* is restricted to calcisiltites and shales (Figure 10) and probably was a "nestler" in some of the larger spaces on and between zoaria. *Zygospira* is extremely abundant and is found in all lithologies where bryozoans occur (Figure 10); it probably nestled in small "nooks and crannies" on and between bryozoans (Figure 7D), but was also capable of attachment to more open areas on crinoid stems and zoaria. Like *Zygospira*, *Rafinesquina* was found in all lithologies (Figure 10) but is relatively uncommon. *Rafinesquina* seems to have preferred open areas on unstable bottoms and provided the initial solid substrates for bryozoan colonization in such areas. Such substrates would have been especially important for the initial development of the bushy ramose colonies previously described. Once the bryozoans became firmly established, however, space was no longer available for *Rafinesquina*, and its numbers declined.

*Rafinesquina* is quite common in the Tanglewood and similar lithologies in the Lexington, and it is possible that the brachiopod may have been an important successional agent in Tanglewood-like lithologies (wavy-bedded calcirudite lithofacies). Ultimate exclusion of *Rafinesquina* may have been the final stage in the succession.

Bioturbation is generally not very apparent in the Sulphur Well. This, we believe, is related to sedimentation rates and the abundance of bryozoans.

In the wavy-bedded calcirudite lithofacies, bioturbation is extremely rare because of the frequency of burial by migrating sand sheets. Although bioturbation is more common in the lenticular calcarenite lithofacies where sedimentation rates were lower, it is still not as common as might be expected. Apparently, the presence of bryozoans, veneering nearly every substrate, formed inpenetrable barriers for bioturbators seeking soft sediments below. The only common trace fossils are *Trypanites* - like borings (Figures 9A and 9B), which may represent phoronid tunnel burrows (Voigt, 1975), and "zigzag" borings of unknown affinity (Figure 9C) which were able to penetrate the massive zoaria. The platy, foliaceous zoaria, on the other hand, apparently were too thin for borers, for they are rarely found bored.

Compared with other shallow, open-marine members of the Lexington Limestone, the species diversity of the Sulphur Well is moderate to low, and most of these species are bryozoans. We believe that this pattern reflects the occupation of most surface habitats by bryozoans and the resulting denial of access to infaunal habitats. This meant that for non-bryozoan fauna to be viable in Sulphur Well environments, they had to be capable of living on irregular or branching zoarial surfaces; this necessarily restricted attendant fauna to a few nestling brachiopods, a few zoarial borers, and a few crinoids able to encrust zoaria.

The overwhelming dominance of bryozoans in the Sulphur Well may reflect the ability of bryozoans to benefit from episodic storms relative to other competitive organisms. The ability of some elevated, rigid forms to withstand partial burial, the possibility of tumbleweed-like transportation, the likelihood of colony proliferation by fragmentation, and the probable rapid growth rates, afforded bryozoans the advantages of immediacy and rapidity in colonizing new sand-sheet substrates. Once the bryozoans became established, it appears that they effectively excluded all other epifauna and infauna except those capable of living on or within the extensive zoaria.

Taphonomically, the fauna of the Sulphur Well largely represents an autochthonous thanatocoenosis. Most fauna appear to have been buried *in situ*. However, in the more calcarenitic lithologies some of the platy zoaria show evidence of transportation, and in the calcirudites nearly all zoaria are fragmented and have been transported (Figures 5B, 5C and 6A). Massive zoaria were prone to being flipped during growth but were not transported into hostile environments, because they continued to grow through accretion by encrustation. Moreover, transportation does not seem to have biased faunal assemblages to a large extent, because patterns of fossil occurrence are obvious and distinct.

## CONCLUSIONS

The Sulphur Well Member of the Lexington Limestone represents a shallow, open-marine transitional environment between a structurally related, high-energy shoal represented by the Tanglewood Member and a quiet, deeper, open-marine environment represented by the Clays Ferry Formation.

The bryozoan-rich lithologies of the Sulphur Well developed only on the protected, leeward side of the Tanglewood shoal. These lithologies originated through bryozoan colonization of sand sheets periodically transported from the shoal during storms. Both lithology and zoarial morphology are clearly related to proxmity to the shoal.

Although periodic storms no doubt devastated bryozoan communities by breakage, burial and transportation, the bryozoans appear to have benefited far more from these effects than other invertebrates. The storms destroyed overcrowded, successionally stable communities, created new firm substrates and aided in the formation and disperal of reproductive units through fragmentation and possible tumbleweed-like transportation. These strategies greatly increased the ecologic fitness of bryozoans in these harsh, transitional environments and afforded them immediacy and and rapidity in colonizing newly created substrates. As a result, the bryozoans were able to exclude other epifauna and infauna, except those able to live on or within zoaria.

The Sulphur Well is among the earliest of bryozoan biostromes, and the diversity of zoarial forms, apparent reproductive modes, and genera seems amazing for a group only a few tens of millions of years after its appearance in the fossil record. Much work on the Sulphur Well remains to be done. Taxonomic studies of the bryozoans are needed, as are studies to discern the relationships that governed interactions between bryozoan species and substrate preference. The possible vertical and lateral succession of bryozoan communities mentioned by McFarlan (1938) and the reasons for it also are areas that require greater scrutiny.

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# LIMESTONE XENOLITHS AND SECONDARY MINERALIZATION IN AN ANALCIME-RICH IGNEOUS DIKE IN HIGHLAND COUNTY, VIRGINIA

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## ABSTRACT

**A dark gray aphanitic igneous rock, occurring as a dike near Monterey, Virginia, contains limestone xenoliths up to 6 cm across. The xenoliths are surrounded by metamorphic halos composed primarily of melilite, magnetite, and perovskite. Subsequent alteration of the dike has converted large portions of the apparently original basalt to an analcime-rich rock (similar to analcimite), and has deposited aragonite, calcite, phillipsite, thomsonite, and pyrite as crusts on cross-cutting fractures. Furthermore, the xenoliths, especially at their peripheries, have been partially altered to tobermorite, ettringite, aragonite, thaumasite, hercynite, and coarser calcite. The rocks and minerals and their relationships are described.**

## INTRODUCTION

Although the igneous dike which is the focus of this study was recognized and mapped over 85 years ago, its nature and complexity have been overlooked until recently. The dike, originally reported as basalt, contains large portions rich in analcime (possibly secondary), and instead of having large amygdaloidal cavities, as earlier described, these surface cavities were formed by the weathering out of limestone xenoliths. The purpose of this paper is to describe the unusual primary, metamorphic, and secondary minerals associated with this igneous body, and to make observations on their genesis. The discovery of tobermorite and aragonite in the dike (Freeland, 1981) initially drew the authors' special attention to the deposit.

## GEOGRAPHIC AND GEOLOGIC SETTING

The site of this study is in western Virginia, about halfway between Monterey and Hightown in Highland County. The igneous dike, which strikes east-west, crosses State Road 637 (Dug Bank Road) a little over a quarter of a mile north of its intersection with U.S. Highway 250, and lies between the lower northwestern flank of Monterey Mountain and Crab Bottom Valley.

Most of the observations reported here were made at a fresh exposure where Road 637 intersects the western portion of the 0.4 mile dike. The dike here is approximately 60 feet wide. Although earlier considered to be Triassic in age, more recent studies by Fullagar and Bottino (1969) indicate that the igneous intrusions in the county are Eocene (average 47 million years). The dike has intruded carbonate rocks of the Beekmantown Formation (Lower Ordovician) and runs eastward into the Martinsburg Formation (Middle and Upper Ordovician). At the road cut the country rock is a limestone containing minor dolomite mineral. The dike has appeared on maps drawn by Darton (1899), Dennis (1934), Parrott (1948,) and Johnson and others (1971), where it has been designated basalt, with an additional comment by Johnson and others (1971) that the eastern end may be andesite or felsite.

## COMMENTS ON THE IGNEOUS ROCK

The dike rock, where it occurs as unweathered material at the road cut, is aphanitic and dark gray. Contrary to earlier reports, it is not typical basalt because it contains considerable analcime. Because of its cryptocrystalline nature, this component is not visible under the microscope, but is easily detected by X-ray diffraction analyses.

The petrographic microscope shows the rock is composed of numerous microphenocrysts of augite with subordinate magnetite and olivine in an abundant, essentially amorphous, groundmass. The augite occurs as clusters of aggregated crystals (like glomeroporphyry), up to 0.9 mm across, and as single crystals 0.5 mm long. This mineral is colorless to pale pinkish-brown, and with crossed nicols shows frequent contact twins and, because of composition variability, an hourglass zonation of interference colors. Magnetite is common and occurs as euhedral to subhedral octahedrons that measure about 0.1 mm across. Frequently these are incorporated within the augite clusters. Although common, olivine is slightly less abundant than magnetite. It occurs as clear, euhedral to subhedral crystals, which usually measure less than 0.1 mm across. These cyrstals commonly show borders stained black and many of them are crisscrossed with very narrow crossfiber alteration veins, possibly serpentine.

The groundmass in which the phenocrysts occur is dark gray. Crossed nicols show that it is composed of amorphous areas and exceedingly small grains with a first order gray interference color. Portions of the matrix exhibit tiny lathlike crystal outlines (approximately 0.1 mm long and less than 0.02 mm wide), resembling plagioclase plates, but under crossed nicols these too are now essentially isotropic, multigranular, and apparently represent altered plagioclase. Because the composition of the groundmass could not be determined optically, over a dozen samples from the road cut were studied by X-ray diffraction analysis (film techniques). The fine-grained matrix is composed primarily of analcime with subordinate nepheline. Two samples which did not contain analcime show excellent diffraction patterns for plagioclase. This suggests that the analcime might have been derived from plagioclase through later alteration, and that some portions of the dike presumably have been unaffected.

Although this study deals mainly with unweathered samples from the road cut near the western end of the exposed dike, rock samples also were collected at intervals along the strike of the outcrop, especially going eastward from the cut. For most of these analcime is also a major component, however, in specimens collected at the easternmost end plagioclase is more prevalent than analcime. About halfway along the strike a coarse phase of the rock contains phillipsite, augite, and plagioclase. At this exposure there are also small amygdaloidal cavities (less that 1 cm agross) that are lined with pseudo-dodechedral phillipsite crystals and that also contain vitreous fibers of thomsonite and rare white, earthy endellite.

Thin sections were made of specimens known to contain plagioclase from both extremities as well as the middle of the dike trend. An attempt to determine the composition of the plagioclase was made by using both the Michel-Lévy and the extinction of microlite methods, but the results were somewhat inconclusive because of the extremely small size of the feldspar laths. A common extinction angle between about 30° and 33°, especially for samples from the eastern end of the dike, indicates sodian labradorite.

Apparently the original igneous rock was an olivine-bearing basalt whose labradorite was sodium-rich and close to andesine. The large portions of the dike that now contain analcime might correctly be named analcimite, an extrusive (or hypabyssal) rock consisting mainly of analcime and pyroxene (ususlly titanian augite), and in which olivine, feldspathoids (e.g., nepheline), plagioclase, and opaque oxides, etc. may be present. Some petrologists who want to reserve the name exclusively for extrusive rocks (and not for hypabyssal bodies like dikes), prefer the name baldite, which essentially has the same composition.

## THE CARBONATE XENOLITHS

Although limestone xenoliths are common in the analcime-rich dike, they went unnoticed for many years, having been interpreted as amygdules filled with calcite, analcime, and "complex aggregates of a number of minerals, some of a clayey nature" (Johnson and others, 1971). At old exposures the xenoliths are weathered out and resemble typical amygdules. On freshly broken rocks, however, one can observe abundant rounded and elliptical cross sections of gray limestone measuring up to 6 cm across (Figure 1*A*). The limestone varies from fine-grained to saccharoidal, and each xenolith is surrounded by a well-defined contact metamorphic zone within the igneous matrx rock. In many instances original small xenoliths have been completely consumed by metamorphic reactions and now consist of orbs composed of the same materials as the contact zones (Figure 1*B*). X-ray diffraction analyses of the xenoliths show they are composed of calcite; however, studies of insoluble residues from these revealed significant melilite. The presence of this mineral, and the often increased grain size of the limestone neighboring the contact, indicate that the limestone itself has undergone recrystallization. Perhaps the melilite here was derived from reactions between original clay (kaolinite) and calcite in the protolith limestone.

## CONTACT METAMORPHIC ZONES AROUND XENOLITHS

Halos of contact metamorphic material surround each of the original xenoliths, or what now are remnants of original xenoliths. Generally these metamorphic zones measure less than 1 cm thick; the small orbs of completely consumed xenoliths are usually less than 2 cm across. The metamorphic material is yellowish gray and contains numerous black spots (0.25 mm across) which, under the microscope, resemble leopard skin. An X-ray diffraction study showed that these zones are composed of melilite and magnetite, with lesser quantities of perovskite.

Thin sections under the polarizing microscope showed an abrupt change between the analcime-rich host rock and the contact zone. In this zone the materials are essentially cryptocrystalline, and with the exception of some of the magnetite, there are no well-defined remnants of the minerals of the host rock. The black, leopardlike spots apparently are pseudomorphs after a subhedral mineral and now consist of nearly amorphous patches with tiny black opaque grains around their circumferences. Because these measure up to 0.25 mm across there is a possibility they were derived from augite, but this is uncertain. Taljaard (1937), however, observed for some South African melilite basalts that earlier formed pyroxene was resorbed and replaced by magnetite, perovskite, and melilite. The Virginia rock sections also showed small areas of calcite within the contact zone.

An electron probe microanalysis was made of one skeletal magnetite crystal (approximately 0.06 mm across) with its melilite-rich matrix. X-ray scanning images, showing Ca, Si, Fe, and Ti corresponded to the respective minerals and also indicated that Ti (possibly from perovskite) was concentrated peripherally around the magnetite.

Where these contact zones have been subjected to weathering, they are earthy, and X-ray diffraction shows that a smectite clay has formed from the melilite. Also the perovskite is more easily detected in these samples because of the diminished amount of melilite.

## ALTERATION OF THE DIKE

After the intrusion of the igneous dike with its included xenoliths, this body presumably was affected by deuteric alteration. This is a situation where water and other volatiles, possibly derived from the same parent magma as the dike at a late stage in its cooling history, invaded the dike and brought about chemical reactions which produced new minerals, especially

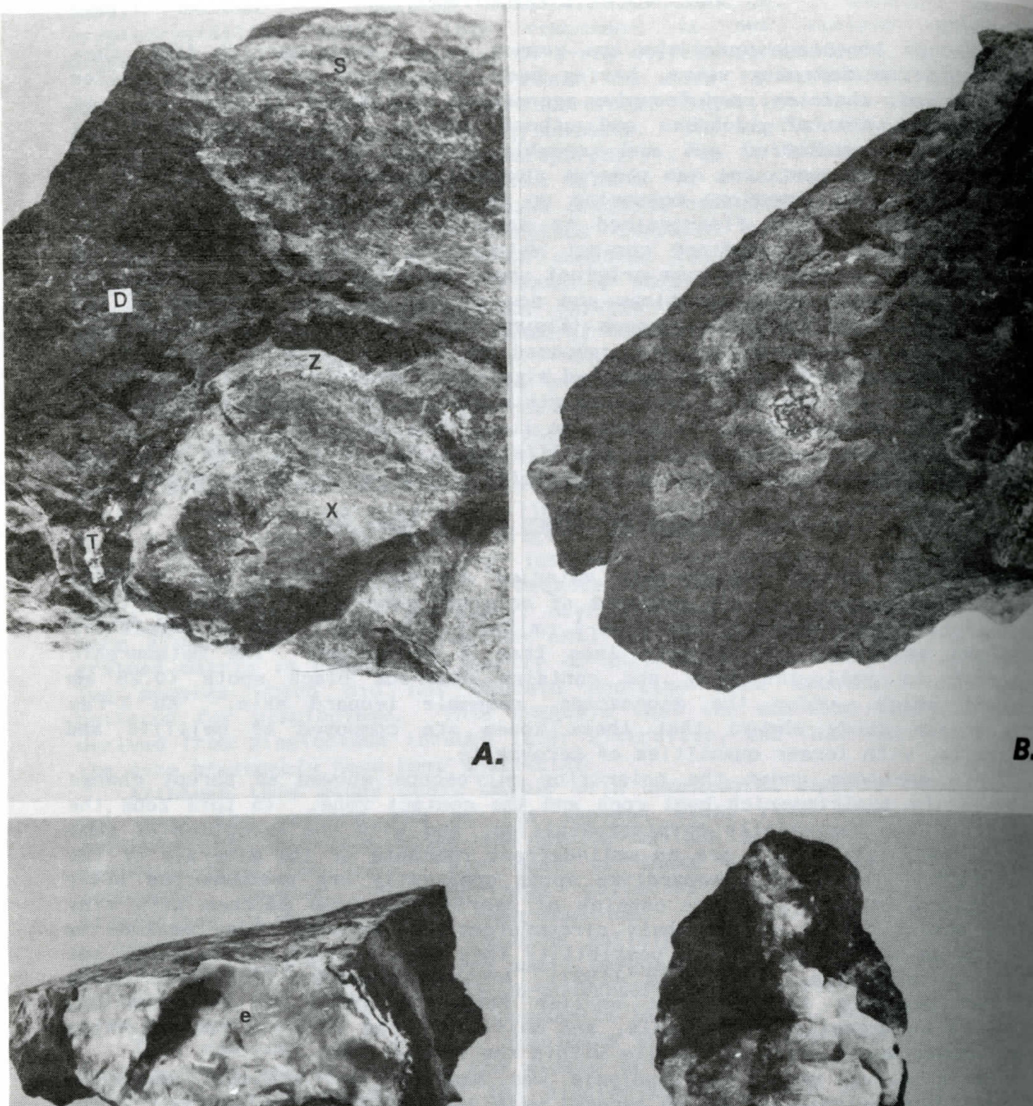


**Figure 1. Limestone Xenoliths...Highland County, Virginia: A. Limestone xenolith (X) showing tobermorite (T) and thin contact metamorphic zone (Z) in dike rock (D). Xenolith 5 cm across. Mineralized seam on fracture surface (S). *B.* Orbs of completely consumed limestone xenoliths (light gray, spotted); one contains tobermorite (white) with black hercynite as dendritic markings. Central orb about 1 cm across. *C.* Tobermorite (white, porcelainoid) with ettringite (e) (prismatic, vitreous). Specimen 4.5 cm long. *D.* Tobermorite (white, porcelainoid) with small radial aragonite (gray) needles. Specimen 3.2 long.**

minerals containing water and sulfate at this deposit. If the analcime, a hydrated mineral, in the matrix of the dike was derived from an earlier plagioclase matrix, as suggested above, this alteration probably occurred at this time. In the presentation that follows, the secondary minerals are divided into two groups, first, those in fractures cutting through the dike, and, second, those occurring within the xenoliths. All the mineral determinations were verified by X-ray diffraction analyses.

## SECONDARY MINERALS ON ROCK FRACTURES

Although the minerals occurring on fractures crisscrossing the analcime-rich rock are very fine grained, and appear as monotonous white crusts, several species were identified through careful studies.

Aragonite, attached to the igneous matrix, occurs on fracture surfaces as vitreous, clear to grayish-white, radiating, sunburst structures, with diameters up to 2 mm. These radiating groups are closely spaced and are very numerous on some surfaces. The mineral often grades into calcite where grains of the latter mineral have replaced it to form clearly visible, but rough, pseudomorphic replacements. Similar aragonite has been observed in cavities and on fracture surfaces in basalt in California (Neuerburg, 1951) and Hungary (Jugovics, 1941).

Calcite also is abundant in fractures as gray, greasy to vitreous crusts with irregular surfaces. Very small subhedral calcite crystals (some nearly pseudocubical in appearance) are occasionally observed on these crusts; the crusts grade into aragonite. On other surfaces calcite occurs a very thin chalky white crusts covering large areas.

Phillipsite, as thin, vitreous to greasy, snow white to gray crusts, overlies aragonite and calcite crusts. The mineral may cover areas measuring several square centimeters. The white crusts seem mottled by gray zones, which are simply thinner areas of the mineral. More rarely phillipsite occurs as tiny (less than a fraction of a mm across), clear crystals with rectangular faces.

Thomsonite, which closely resembles phillipsite here, is less common. It occurs as thin, white to gray, vitreous crusts usually superimposed on phillipsite The mineral was distinguished from phillipsite only through X-ray diffraction analyses.

Tobermorite, as small areas rarely over 2 mm across, occurs as distinctly white, chalky to silky material showing a radial fibrous structure with slight concentric banding. It fills the interstices between calcite and other grains. Although the X-ray data for most of this material indicates typical tobermorite, some data exhibit slight variations, and may represent altered tobermorite that approaches related minerals with slightly different structures and water contents, e.g., plombierite.

Pyrite crystals with cubic habits rarely occur with the other secondary minerals on fracture surfaces. These are a fraction of a millimeter across and are very scarce on most specimens.

## SECONDARY MINERALS ASSOCIATED WITH XENOLITHS

The xenoliths frequently are replaced by secondary minerals or have zones of replacement especially at their contact with the well-defined metamorphic rims that surround them (Figure 1A).

Tobermorite is the most abundant of these minerals, located especially at the peripheries of altered xenoliths. It occurs as white, fine-grained, porcelainoid masses (Figure 1C). Bluish-gray tobermorite also occurs in botryoidal form at the contacts between xenoliths and the metamorphic halos, and it is arranged so convex botryoidal surfaces point inward toward the xenolith center. In appearance this variety closely resembles chalcedony. Another rare form of tobermorite is found as velvety crusts, on massive tobermorite, and consists of white silky fibers perpendicular to the crust surfaces.

Ettringite crystals, up to 4 mm wide and over 5 mm long, are embedded in some of the tobermorite (Figure 1*C*). These colorless, hexagonal crystals are clear, vitreous, and show excellent cleavages at 120°. Some specimens which are now chalky to pearly white are calcite pseudomorphs after ettringite. Goniometric studies of one of these (2 mm by 5 mm) show all three planes of the prismatic cleavage despite its altered and somewhat skeletal nature.

Aragonite prisms are attached to the walls of contact between the xenoliths and the metamorphic halos and have a delicate sunburst radial arrangement pointing into, and embedded within, the tobermorite and other xenolith minerals (Figure 1*D*). The individual prisms range from thin fibers to needles about 0.25 mm wide and over 5 mm long. They are yellow-gray and greasy to vitreous.

Thaumasite, as white masses, occurs in contact with tobermorite. It is easily differentiated because it has a coarser texture and has a somewhat pearly talclike appearance. Where thaumasite surfaces are exposed to open cavities within xenoliths, tiny, acicular crystals occur.

Hercynite is the major component of very thin black sinuous to dendritic markings within the porcelainoid tobermorite (Figure 1*B*). X-ray diffraction analyses yielded a = 8.14 ± 0.01 Å, which supports hercynite and excludes other related spinels.

Calcite, as fairly large grayish grains and rarely as euhedral crystals, occurs with the rarer minerals, however, it generally is more abundant in zones away from these minerals and is closely associated with recrystallized xenolith limestone.

**Table 1. Minerals in the Deposit.**

Contact Metamorphic Minerals around Xenoliths

| | |
|---|---|
| Magnetite, | $FeFe_2O_4$ * |
| Melilite, | $(Na,Ca)_2(Mg,Fe,Al,Si)_3O_7$ |
| Perovskite, | $CaTiO_3$ |

Secondary Minerals

| | | Dike rock | Seams on fractures | With xenoliths |
|---|---|---|---|---|
| Analcime, | $NaAlSi_2O_6 \cdot H_2O$ * | x | | |
| Phillipsite, | $(K,Na,Ca)_{1-2}(Si,Al)_8O_{16} \cdot 6H_2O$ | | x | |
| Pyrite, | $FeS_2$ | | x | |
| Thomsonite, | $NaCa_2Al_5Si_5O_{20} \cdot 6H_2O$ | | x | |
| Tobermorite, | $Ca_5Si_6O_{16}(OH)_2 \cdot 4H_2O$ | | x | x |
| Aragonite, | $CaCO_3$ | | x | x |
| Calcite, | $CaCO_3$ | | x | x |
| Ettringite, | $Ca_6Al_2(SO_4)_3(OH)_{12} \cdot 26H_2O$ | | | x |
| Hercynite, | $FeAl_2O_4$ | | | x |
| Thaumasite, | $Ca_3Si(CO_3)(SO_4)(OH)_6 \cdot 12H_2O$ | | | x |

* All formulas from Fleischer, Glossary of Mineral Species, 1983.

## REMARKS ON PARAGENESIS

It appears that the following events occurred during the history of the igneous dike:

a. The dike, possibly of basaltic composition and containing limestone xenoliths, was emplaced into Ordovician carbonate rocks during the Eocene.
b. The xenoliths underwent high temperature, low pressure sanidinite metamorphism to form halos composed primarily of melilite, magnetite, and perovskite.
c. Deuteric alteration, from volatiles rich in water and sulfur,

brought about the formation of new minerals (Table 1).

1. The plagioclase in the igneous rock was altered to analcime, although remnants of the feldspar are present in some portions of the dike.
2. Interactions between volatiles and igneous rock fracture surfaces (and rare small amygdules) yielded phillipsite, thomsonite, pyrite, as well as aragonite, calcite and rare tobermorite.
3. The already recrystallized limestone xenoliths were partly or wholly altered to form calcium-rich tobermorite, ettringite, thaumasite, and aragonite.

## ACKNOWLEDGMENTS

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# STRUCTURAL AND METAMORPHIC EVOLUTION OF A PORTION OF THE BLUE RIDGE IN MARYLAND

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## ABSTRACT

**Four phases of folding and at least one metamorphic event were identified in a portion of the eastern limb of the Blue Ridge anticlinorium in Maryland. The youngest three fold phases post-date greenschist metamorphism and are related to formation of the anticlinorium during the Alleghanian orogeny. The earliest phase of folding and associated metamorphism pre-date formation of the anticlinorium and are belived to be pre-Alleghanian. Thus, it can be argued that rocks in this portion of the Blue Ridge have been affected by two separate Paleozoic orogenic events, not just the Alleghanian as is generally accepted.**

## INTRODUCTION

Thought concerning the Blue Ridge has evolved considerably over the past 20 years. Whereas the Blue Ridge in the central Appalachians was thought to be autochthonous and the cause of the deformation in the Valley and Ridge (Cloos, 1947, 1971, 1972), it is now generally accepted that the Blue Ridge is allochthonous and a part of the same deformation plan that includes the Valley and Ridge, as well as portions of the Piedmont (Root, 1970, 1973; Harris, 1979; Harris and Bayer, 1979; Harris and others, 1981; Harris and others, 1982; Cook and others, 1982). The rootless character of the Blue Ridge in the southern Appalachians has long been recognized (Bryant and Reed, 1970), but it was not until the work of Root (1970, 1973) and Harris (1979) that it became apparent that the allochthonous nature of the Blue Ridge extends northward to its terminus in Pennsylvania.

Recognition of the allochthonous nature of the Blue Ridge in the central Appalachians has important implications for the geological evolution of this region. Earlier workers regarded the Blue Ridge as a shear fold involving basement and cover rocks acting as a single unit (Cloos, 1947, 1971, 1972). Now the Blue Ridge is thought to be a thick-skinned analogue of the rootless anticlines of the Valley and Ridge (Harris, 1979).

Although the present location of the central Appalachian Blue Ridge and much of the deformation is thought to be the result of Alleghanian thrusting, structural and metamorphic features in a portion of the Blue Ridge in Maryland suggest that some of the rocks exposed there may have also been affected by an earlier, possibly pre-Alleghanian event.

### Geologic Setting

In Maryland, the Blue Ridge consists of a northeast-plunging, asymmetric anticlinorium, locally overturned to the northwest (Figure 1). The oldest rocks exposed in the core of the structure are Grenville granodiorite and granite gneiss. These are overlain unconformably by the Late Precambrian Swift Run and Catoctin Formations (King, 1951). The Swift Run consists of phyllites and impure quartzites and is present only locally. The Catoctin is comprised of greenstone with subordinate metarhyolite and metasediments. Overlying the Catoctin are the clastic rocks of the Lower Cambrian Chilhowee Group consisting of the Loudoun, Weverton, Harpers, and Antietam Formations, from oldest to youngest. The contact between the Catoctin and Chilhowee along the Blue Ridge has been interpreted by Nickelsen (1956) and Cloos (1951) to be conformable and by King (1951) to be unconformable. Whitaker (1955) studied the structure and stratigraphy along Catoctin mountain,

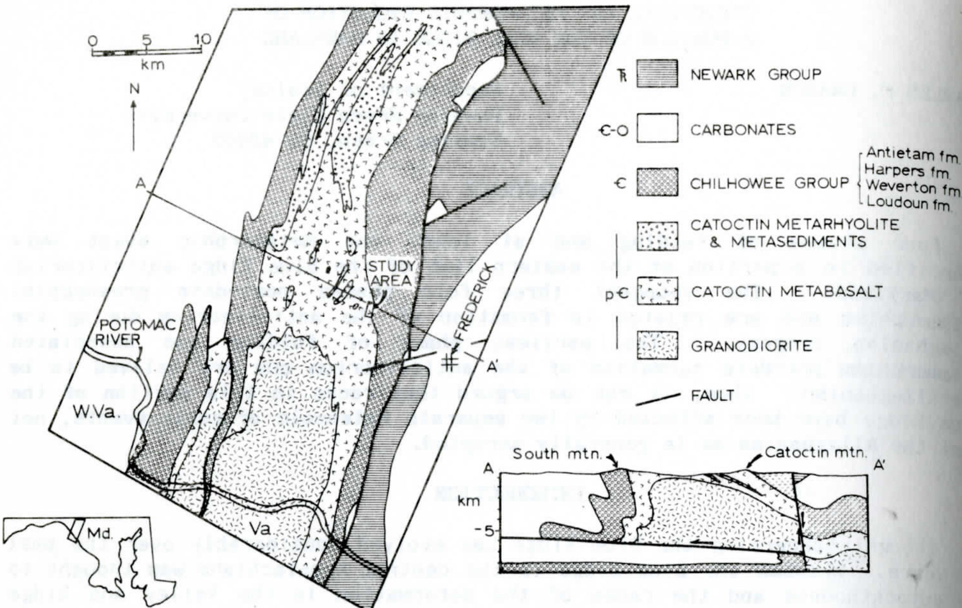


**Figure 1. Generalized geologic map from Cleaves and others (1968) and cross section of the Blue Ridge in Maryland and surrounding vicinity. Inset map shows location of geologic map. Cross section is modified from Perry and de Witt (1977).**

including the area described in this paper, and concluded that the Catoctin grades upward into phyllites of the basal Chilhowee Group without any obvious break.

The area described in this paper is along Catoctin Mountain, 8 km west of Frederick, Maryland (Figure 1). Exposures were examined along the crest and west flank of Catoctin mountain from High Knob in Gambrill State Park, northward for a distance of about 1.5 km. Structurally, the area is on the eastern, upright limb of the anticlinorium. Volcanically-derived phyllites in the uppermost portion of the Catoctin Formation are well exposed throughout the study area. In this area, the contact with the basal unit in the overlying Chilhowee Group, the Loudoun Formation, is gradational. The mineralogy and intrusive features suggest that the phyllites examined belong to the uppermost portion of the Catoctin. Most of the phyllites contain relatively high percentages of chlorite, more typical of Catoctin phyllites. The phyllites are mostly silver to gray-green, not purple as is characteristic of the Loudoun phyllites (Whitaker, 1955). One-centimeter wide, pre-metamorphic mafic dikes, typical of those found elsewhere in the Catoctin, were found in the phyllites suggesting that the host rock has a closer affinity to the volcanic Catoctin than to the sedimentary Loudoun.

The phyllites are ideally suited for a structural study. Due to their relative incompetence, they preserve a more detailed record of the deformational history than is found in the overlying Chilhowee clastics or the more massive portions of the underlying Catoctin Formation. Cleavages, lineations, and other penetrative structural features, common in the phyllites, are sparse or absent in the under-and overlying rocks.

## STRUCTURAL GEOLOGY

The structural evolution of the area was determined from the temporal and spatial relations between the numerous mesocopic and microscopic structural features in the phyllites. From these, four phases of folding were recognized. This is in contrast with previous studies of this and nearby areas in the Blue Ridge which recoginze only two phases of folding (Whitaker, 1955; Nickelsen, 1956; Fauth, 1968).

### $D_1$

The oldest structures in the area are isoclinal folds in the compositional layering, $S_0$. These folds are difficult to recognize due to the subsequent effects of $D_2$. An axial planar schistosity, $S_1$, defined by the preferred orientation of muscovite, chlorite, and chloritoid, is present in most outcrops. This phase has not been described in previous studies of this part of the Blue Ridge.

### $D_2$

Second phase structures dominate the mesoscopic and microscopic fabrics of the phyllites. $F_2$ folds are generally tight, but vary from isoclinal to moderately open. All have an asymmetry that indicates an east-over-west movement sense. $S_2$ is a crenulation cleavage that is axial planar to $F_2$ folds. Where intensely developed, it partly or totally obliterates $S_1$ (Figures 2 and 3) and is better termed as schistosity. In many samples, all that remains of $S_1$ are small lenticular domains of chlorite and muscovite with their 001 cleavage at high angles to the enclosing $S_2$ surfaces.

$D_2$ structures are temporally and spatially associated with the Blue Ridge anticlinorium. $F_2$ folds are coaxial with the anticlinorium and have the correct asymmetry for their location on the upright limb. $S_2$ is parallel to the axial plane of the anticlinorium. $D_2$ of this study correlates with the first phase in earlier studies (Whitaker, 1955; Nickelsen, 1956; and Fauth, 1968).

### $D_3$

Third phase structures consist of open or kink folds with an axial planar crenulation cleavage, $S_3$. These structures are not everywhere developed, but where present, they clearly deform $D_2$ structures (Figure 2). This phase correlates with the second phase of Whitaker (1955), Nickelsen (1956), and Fauth (1968).

### $D_4$

The youngest structures are open folds, or more rarely, kink folds with a poorly developed axial planar crenulation cleavage, $S_4$. These structures were found only in a few locations in the study area. $D_4$ has not been recognized in previous studies.

## Orientation Data

Plots of the structual features discussed in this paper are shown in Figure 4. Most plots were not contoured to avoid the unrealistically high contour levels inherent with contouring data sets of less than 50 points.

The first three fold phases are coaxial and plunge gently to the northeast or southwest, parallel to the trend of the anticlinorium. Although the range of fold axis orientations overlaps, the first three phases can be differentiated on the basis of their average S-surface orientation (Figure 4). $S_1$ dips moderately to the northwest, $S_2$ dips gently to the southeast, and $S_3$ dips steeply to the northwest. Fourth phase structures are oriented almost at right angles to earlier structures: $F_4$ folds plunge gently SE or NW and $S_4$ dips steeply NE or SW.

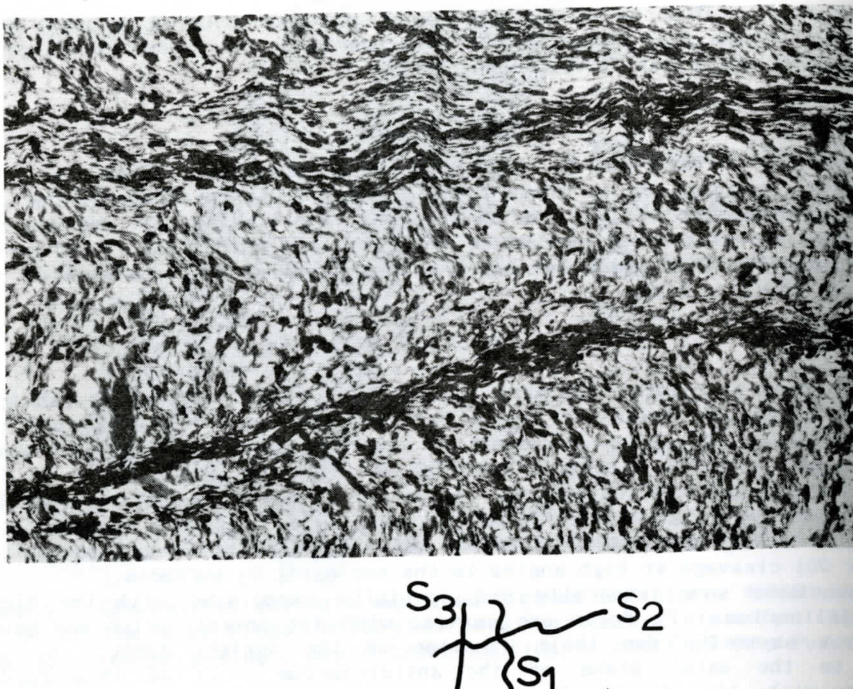


Figure 2. Photomicrograph showing $S_1$, $S_2$, and $S_3$. Note local obliteration of $S_1$ by $S_2$ and relative abundance of magnetite in zone where $S_2$ is developed. Width of view is 1.0 mm.

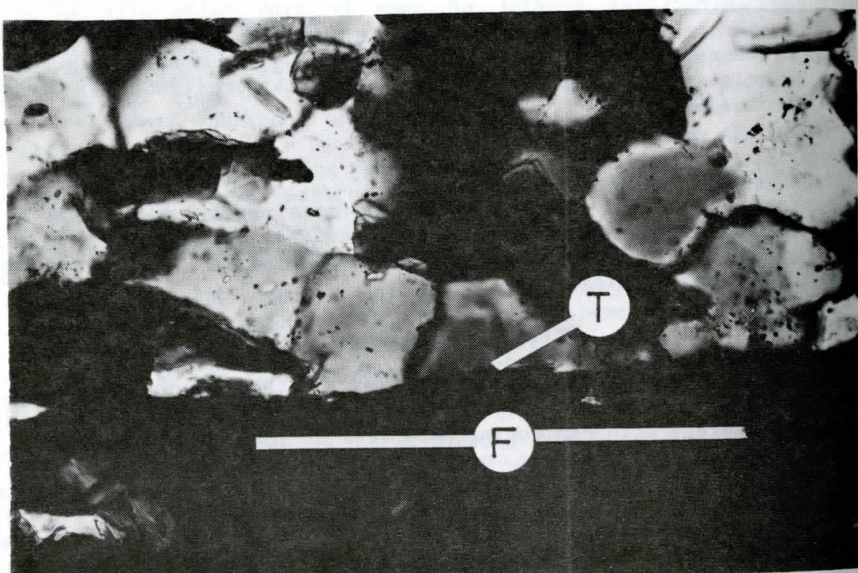


Figure 3. Photomicrograph showing features resulting from pressure solution. F - $S_2$ folia composed of dark concentration of phyllosilicates and opaques. T - quartz grains truncated by folia. Width of view is 0.2 mm.

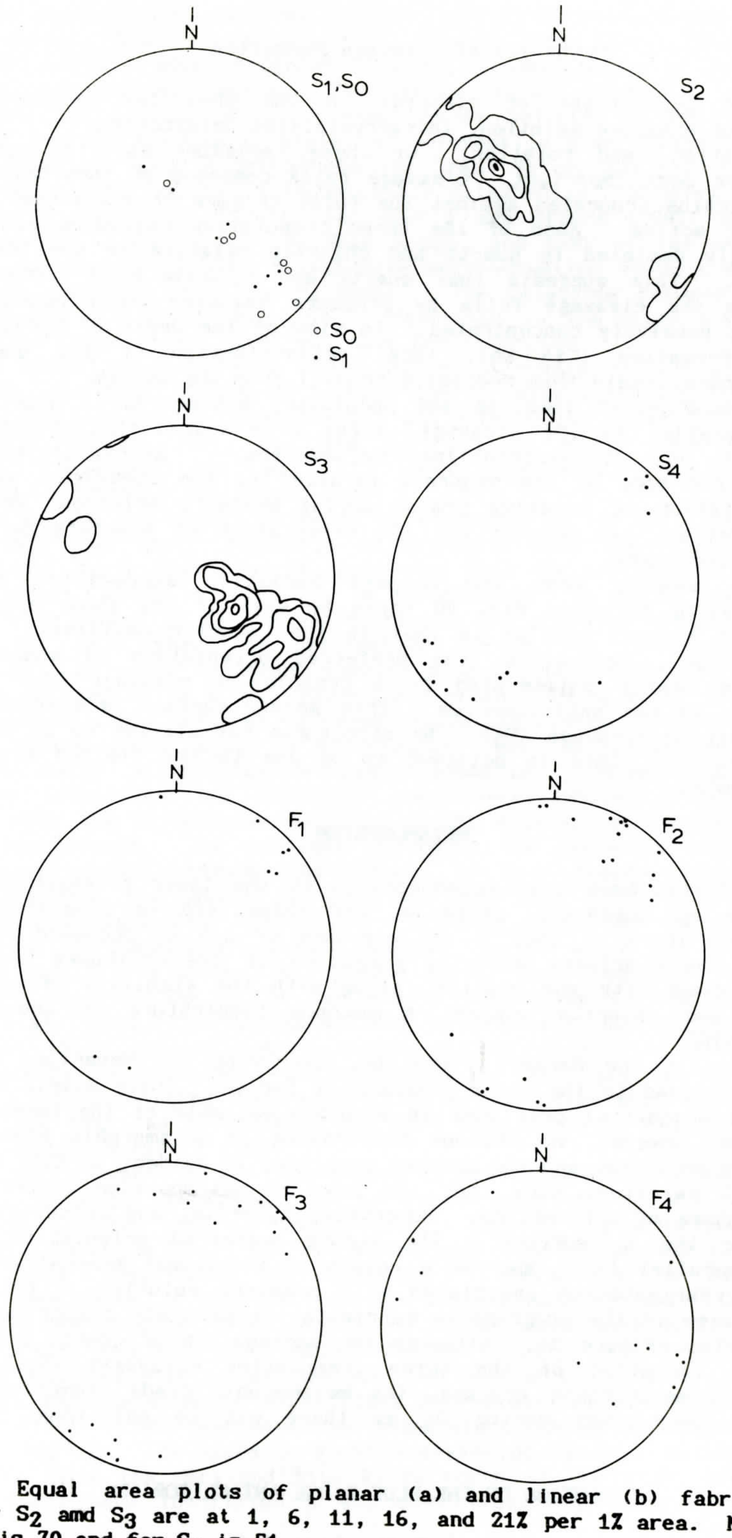


Figure 4. Equal area plots of planar (a) and linear (b) fabric data. Contours for $S_2$ amd $S_3$ are at 1, 6, 11, 16, and 21% per 1% area. Number of data for $S_2$ is 70 and for $S_3$ is 51.

## Mechanics of Cleavage Formation

The four generations of cleavage in the phyllites developed by a combination of pressure solution, intracrystalline deformation, recrystallization, and rotation. Of these, pressure solution appears to have been the most important. Cleavage folia composed of insoluble residue and quartz grains truncated against the folia (Figure 3) argue that pressure solution was active. Each of the three crenulation cleavages consists of folia strongly depleted in quartz and chlorite relative to the intervening microlithons. This suggests that quartz and chlorite were preferentially removed from the cleavage folia by pressure solution while muscovite and opaques were passively concentrated. In view of the depth of burial at the time of deformation (~14 Kb), local redistribution of the quartz and chlorite is more likely than wholesale removal from the system.

Sparse deformation lamellae and undulatory extinction in quartz grains elongated parallel to the cleavage, along with kinked phyllosilicates are the product of intracrystalline deformation. Recrystallization is responsible for some of the minerals parallel to the cleavages, especially in $S_1$. Rotation of existing grains during pressure solution, as well as recrystallization, can account for the orientation of minerals parallel to the younger cleavages.

Phyllite samples from the Harpers Formation exposed in a recent excavation along Maryland Rte. 40 three km east of the study area show a deformational history similar to that in the Catoctin phyllites. An early schistosity ($S_1$), defined by the preferred orientation of muscovite and chlorite, is nearly obliterated by a crenulation cleavage ($S_2$) which is axial planar to the anticlinorium. This second surface is crenulated by a northwest-dipping cleavage ($S_3$). No structures correlative to $D_4$ were found in this outcrop. This is believed to be due to the limited size of the outcrop examined.

## METAMORPHISM

The phyllites have been metamorphosed to the lower greenschist facies. The presence of muscovite, chlorite, and chloritoid is consistent with a minimum temperature of 350° C and pressure of 3.5 Kb proposed by Elliott (1973) for the overlying Weverton Formation of the Chilhowee Group. The absence of staurolite and biotite, along with the stability of chloritoid, muscovite, and chlorite suggest a maximum temperature of about 420° C (Winkler, 1976).

Phyllites in the Harpers Formation, overlying the Weverton, have also been metamorphosed to the lower greenschist facies. Their simple mineralogy (quartz and muscovite) precludes an accurate estimate of the temperature of metamorphism; however, no obvious discordance in metamorphic grade between the Catoctin phyllites and the Harpers phyllites is evident in this area.

Textural relations show that the peak of metamorphism occurred during the first phase of deformation. Muscovite, chlorite, and chloritoid all lie parallel to the $S_1$ surface. The strong preferred orientation of layer silicates parallel to $S_2$ may be a result of additional mineral growth, but mechanical reorientation associated with pressure solution is perhaps more likely in view of the progressive reorientation of platy grains near the $S_2$ cleavage folia (Figure 3). Although the segregation of quartz and chlorite during the formation of the three crenulation cleavages ($S_{2-4}$) can be considered a metamorphic process, the metamorphic grade (temperature) must have been lower than during $D_1$ as there was no additional growth of chloritoid.

## EVOLUTION OF THE BLUE RIDGE ANTICLINORIUM

The generally accepted view of the Blue Ridge anticlinorium in Maryland is that the metamorphism and most of the folding occurred during a single event (Nickelsen, 1956; Freedman, 1967; Fauth, 1968; Root, 1970; Cloos, 1971). Most workers also recognize a minor, younger phase of folding and crenulation cleavage development. The age of the anticlinorium, associated metamorphism, and younger minor deformation is believed to be Alleghanian because the overall structural style, or deformation plan (Cloos, 1971), is continuous from the Blue Ridge across the Great Valley and into the Valley and Ridge where deformation is indisputably Alleghanian. Studies by Harris (1979), Harris and Bayer (1979), and Harris and others (1981) have shown that the present location of the Blue Ridge, and most of the deformation within it, is associated with Alleghanian mega-thrusting above a detachment that extends eastward into the Piedmont. The relations between the four fold phases and one metamorphic event identified in the study area suggest that more than one orogenic event may have been involved.

On the basis of structural style and relationship to the greenschist metamorphism, the four phases of folding can be separated into two events. The $F_1$ isoclinal folds, $S_1$ schistosity, and greenschist metamorphism belong to the first event. $D_2$, $D_3$, and $D_4$, all of which have associated crenulation cleavages and all of which post-date the greenschist metamorphism, constitute the second event.

The age of the two events relative to formation of the anticlinorium can be determined from their geometric relationship with anticlinorium. As stated earlier, $D_2$ is most closely associated with the anticlinorium; hence, the second event, which includes $D_2$, is Alleghanian. $D_3$ and $D_4$ are probably associated with formation of the anticlinorium and are also Alleghanian. $D_2$ would be the main phase of deformation during the formation of the anticlinorium. $D_3$ would represent coaxial progressive deformation following $D_2$. $D_4$ is belived to be a very minor late cross folding event of the type common in many multiply deformed terranes (Hobbs and others, 1976).

If the second event ($D_{2-4}$) is Alleghanian, then the first event ($D_1$) is either very early Alleghanian or pre-Alleghanian. Geiser and Engelder (1983) have identified two distinct, non-coaxial phases of deformation associated with the Alleghanian orogeny in the Appalachian foreland of New York and Pennsylvania. The earliest of these, the Lackawana phase, is as old as Pennsylvanian, whereas the latest, or Main phase, is Permian or younger. The two, non-coaxial deformations described by Dean and Kulander (1977) in southwestern West Virginia are believed by Geiser and Engelder to correlate with the Lackawana and Main phases. Harris and Milici (1977) also mention the possibility of multiple phases during the Alleghanian orogeny in the southern Appalachians. The two events identified in the study area could correlate with the Lackawana and Main phases of the Alleghanian orogeny, but their coaxial nature argues against this. In general, the large difference is structural style between $D_1$ and $D_{2-4}$, along with the fact that $D_1$ is associated with significant metamorphism whereas the others are not, argue that the differences are more than would be expected between phases of a single orogenic event.

If the early event in the study area is pre-Alleghanian, then it could be Acadian, Taconian, Avalonian, or an event unrelated to any recognized tectonic event in the central Appalachians. Several arguments can be made in favor of a Taconian age. (1) $D_1$ is the only one of the four fold phases associated with significant metamorphism. There are no Alleghanian metamorphic age dates in this part of the central Appalachians. Ordovician shales of the Martinsburg Formation, 25 km to the west, were deformed in the Alleghanian, yet they have been metamorphosed to only the lowest anchizone grade. The detrital plagioclase has not lost any of its calcium (Wright and Kreps, 1979) and there is no significant recrystallization (Onasch, 1983). Wright and Platt (1982) estimate the temperature during deformation as no higher than 300° C. (2) Based on Rb-Sr analyses of Catoctin metabasalt and metarhylolite from the Maryland Blue Ridge north of the study area, Nagel and Mose (1984) have identified a 420 m.y. (Taconian) metamorphic event.

(3) To the north in Pennsylvania, development of nappes in the Lebanon Valley took place during Taconic (MacLachlan and Root, 1966) demonstrating that significant deformation occurred during the Taconic in nearby areas. (4) $D_1$ and associated metamorphism cannot be Avalonian because the Lower Cambrian phyllites of the Harpers Formation show a similar metamorphic and deformational history to the phyllites in the study area. (5) An Acadian age for $D_1$ seems unlikely in view of the absence of Acadian dates in or west of the Blue Ridge of the central Appalachians. (6) Assigning $D_1$ to an event unrelated to any of the recognized events in the Appalachians is possible, but the author prefers to interpret these data in the framework of established orogenic events until positive evidence to the contrary is found.

By the process of elimination, suggestive evidence of (1) and (3), and the positive evidence of (2), a Taconian age for $D_1$ and associated greenschist metamorphism is favored. It should be emphasized that a Taconian event did not affect the area presently occupied by the Blue Ridge. The rocks have been transported a minimum of 59 km northwestward during Alleghanian thrusting (Harris, 1979), placing them in the present-day Piedmont at the time they underwent $D_1$ and metamorphism.

The structural and metamorphic relations just described for the eastern limb of the Blue Ridge differ in two ways from those reported by Nickelsen (1956) at South Mountain, along the western limb of the anticlinorium. (1) Only two phases of folds and associated cleavages are present at South mountain. $D_1$ and $D_4$ are absent. (2) On the Western limb, metamorphism was synchronous with the formation of the anticlinorium whereas on the eastern limb, metamorphism predates it. Metamorphic minerals are parallel to the axial plane of the anticlinorium at South mountain, but on the eastern limb, they are cut by the axial planar cleavage. Most of these differences could be explained in light of the higher strains present along the western limb that are associated with formation of the anticlinorium (Mitra, 1976). A pre-anticlinorium phase of deformation on the western limb, correlative to $D_1$ on the eastern limb, could have been totally obliterated during formation of the anticlinorium in the Alleghanian. Metamorphic minerals formed during this early phase would have been rotated into alignment with the axial surface giving the appearance that metamorphism was coeval with the formation of the anticlinorium. This model, although based on negative evidence, is supported by the local obliteration of $S_1$ by $S_2$ on the eastern limb. With the additional strain present on the western limb, $S_1$ might no longer be visible.

Textures in Harpers samples from Harpers Ferry and elsewhere along South mountain examined by this author are consistent with this interpretation. Although the existence of an earlier cleavage cannot be proven conclusively, the dominant cleavage more closely resembles $S_2$ at Catoctin mountain than $S_1$.

## SUMMARY

Analysis of the minor structural features from a portion of the eastern limb of the Blue Ridge in Maryland has revealed four phases of folding and one phase of greenschist metamorphism. Of these, the youngest three fold phases are related to the formation and subsequent modification of the Blue Ridge anticlinorium during the Alleghanian orogeny. The earliest phase of folding and associated greenschist metamorphism predate formation of the anticlinorium and are believed to be Taconian in age. Hence, it can be argued that the Blue Ridge in this portion of the central Appalachians has been affected by two orogenic events, not just one as is generally accepted.

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# THE SOUTH FORK OF THE LICKING RIVER—EASTERN KENTUCKY'S MAJOR LATE TERTIARY RIVER?

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## ABSTRACT

**The South Fork of the Licking River in east-central Kentucky is a relatively small stream. Evidence—much of it indirect in nature—now indicates that, in preglacial late Tertiary time, the South Fork was a much longer stream, drained a larger area, and may have been the trunk stream of the Licking River system.**

**Five principal lines of evidence support this conclusion: (1) The present stream is manifestly underfit, as was its pre-Illinoian (Teays-age) predecessor. (2) Quartz and chert clasts whose bedrock source is well beyond the present drainage area are abundant in stream gravels of the Teays-age valley. (3) A wide, well-defined valley intermediate in elevation between the present upland surface and the Teays-age valley apparently headed far beyond drainage divides. (4) Wavelengths of Teays-age valley meanders are larger than those of the present South Fork and out of proportion to the present drainage area. (5) Geomorphological characteristics of the Kentucky River Gorge suggest it was the channel by which the South Fork's former headwaters could have been captured by the Kentucky River.**

**Probably through glacial damming and drainage disruptions during an early Pleistocene but pre-Teays glaciation, the former headwaters of this ancestral South Fork were captured by a Kentucky River that was actively and rapidly cutting a new course through the Kentucky River Gorge along zones of lithologic weakness created by repeated movement along the Kentucky River fault system.**

## INTRODUCTION

The South Fork, which is the largest tributary of the Licking River of east-central Kentucky, is a relatively small north-flowing stream with a drainage area of only 2,400 $km^2$ (927 $mi^2$). Its head lies a short distance upstream of Shawhan (fig. 1) at the junction of Stoners and Hinkston Creeks, and it joins the main stem of the Licking River at Falmouth.

The walls and floor of the South Fork's valley consist of Ordovician limestone, shale, and siltstone, which are highly variable in their proportions and relative competence. For nearly its entire length, the South Fork is a meandering stream within a meandering valley and has all the general characteristics of a manifestly underfit stream as defined by Dury (1964, p. A6-A9).

In an earlier contribution to the geologic history of part of the Licking River, I proposed (Luft, 1980, p. 7-8) that the South Fork, rather than the Licking itself, had all the attributes of being the trunk stream of pre-Illinoian (Teays) time. There the matter rested, because of serious obstacles to proving that contention. I have since considered and accepted the following supporting evidence in favor of a perhaps considerably longer course of the South Fork in pre-Teays time:

(1) At present the South Fork valley is wider than the Licking River valley upstream from their junction.

(2) Quartz and chert gravel are abundant in Teays-age and later fluvial deposits of the South Fork, though no such bedrock material is presently available within the South Fork's drainage basin.

(3) A wide, pre-Teays, north-sloping valley surface, present in an intermediate position between the upland surface and the Teays-age Parker strath delineated in Luft (1980), suggests that the pre-Teays drainage basin

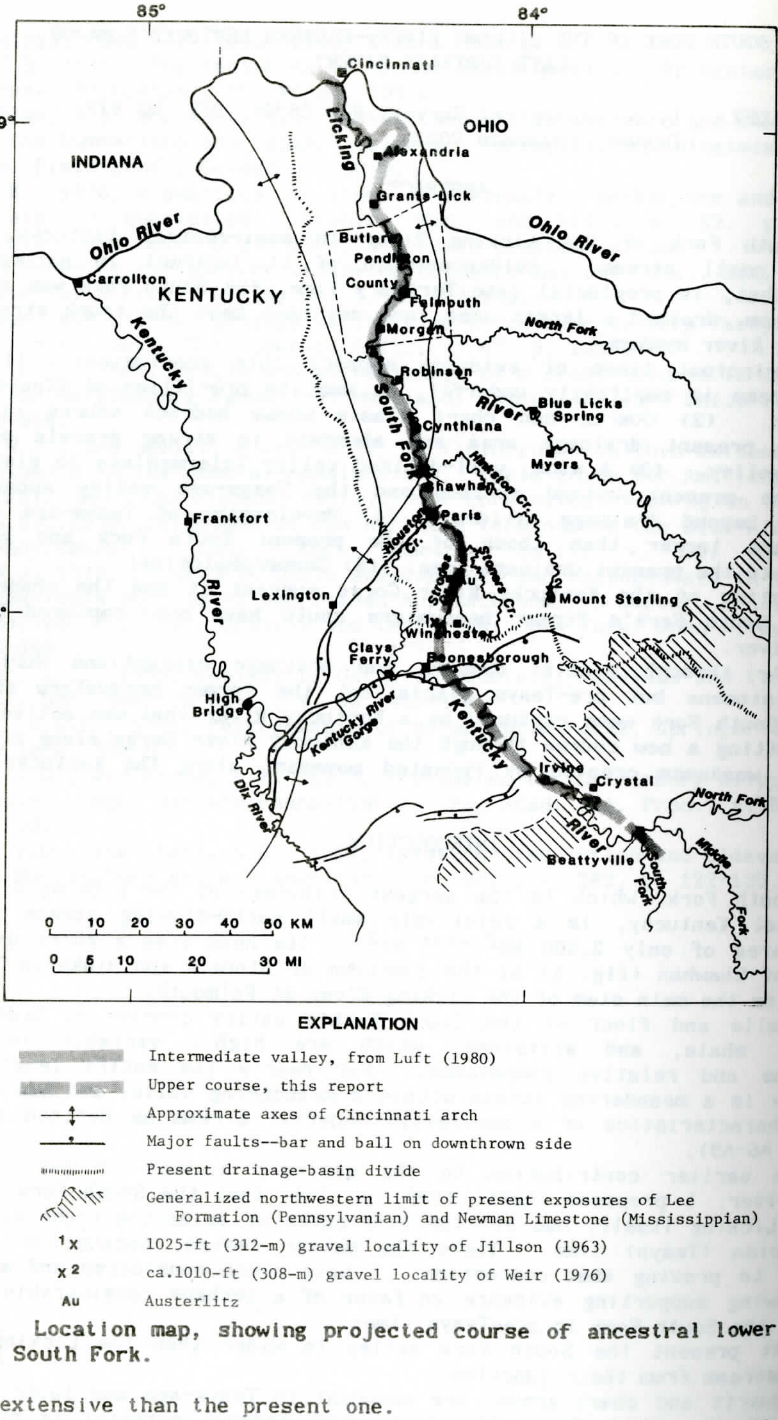


**Figure 1. Location map, showing projected course of ancestral lower Licking River and South Fork.**

was more extensive than the present one.

(4) Meander wavelengths of the Teays-age South Fork are larger than those of valley meanders of the present South Fork and are out of proportion to its present, small drainage area.

(5) The geomorphological characteristics of the Kentucky River Gorge suggest that it was the channel by which the neighboring Kentucky River

captured the headwaters of the ancestral South Fork and reduced the South Fork's drainage area to present proportions.

Item 1 is obvious, from perusal of topographic or geologic quadrangle maps of the region. The close parallelism of the proposed course to the Kentucky River-Dix River trend is noteworthy and will be discussed at the end of this article. The other items of evidence are treated here in greater detail.

## SOURCE OF THE EXOTIC GRAVEL

Since at least Teays time, the lower Licking River has flowed northward upon Ordovician and Silurian bedrock within the study area, but its bedload contains considerable amounts of coarse gravel derived principally from the Pennsylvanian Lee Formation and the Upper Mississippian Newman Limestone. Abundant quartz and chert clasts and lesser amounts of sandstone and coralline limestone clasts are derived from erosion of source rocks exposed to the east of the area of figure 1. The South Fork, which flows northward entirely upon Ordovician strata, also carries abundant similar gravel, albeit of finer size. Sources for large amounts of exotic gravel are lacking within the present confines of the South Fork's drainage basin.

Some suggested sources of this stream gravel can be readily discounted as being, at best, very minor contributors to the total exotic gravel fraction in South Fork Teays-age deposits (Luft, 1980, p. 8). These include (1) reworked or let-down older, residual fluvial gravels, and (2) downstream gravel of the Licking River bedload brought in during drainage reversal.

Residual upland gravel deposits could only have been minor contributors to the South Fork's bedload. Exposures of gravel on surfaces above the levels of Teays-age fluvial deposits are scarce and generally of minimal extent. The exotic gravel contained is less abundant and generally much finer grained than the average for Teays-age deposits.

Drainage reversal along the Licking River system, and its supposed effect upon the gravel bedload, is extremely unlikely and is mentioned here only for the record. Miller (1895) postulated that, as the result of an early glacial ponding in the vicinity of Cincinnati, Ohio, the South Fork became a temporary sluiceway for backed-up waters of the upper Licking River system. The waters supposedly flowed southward into the South Fork and escaped over a low divide into the Kentucky River to the west. Thus, gravel from source areas along the main stem of the Licking River could have been brought into the then valley of the South Fork. However, most tributaries (and paleotributaries) of the South Fork are not barbed, and so this hypothesis of drainage reversal becomes untenable. Furthermore, it does not appear that the high-level fluvial sediments above the South Fork were deposited in Teays (i.e., preglacial) time by other than the normal fluvial processes of a north-flowing stream.

Only one plausible source remains: direct derivation from the Lee Formation and the Newman Limestone. The nearest outcrops of these formations lie at least 20 km from present-day headwaters of the South Fork (figure 1). They are now drained by the Kentucky River and the main stem of the Licking River, but not by the South Fork. It is unlikely that extensive outcrops of these resistant formations receded 20 and more kilometers during the short interval since Teays time. It is therefore postulated that a longer stream than the modern South Fork, at one time, tapped the extensive outcrop areas of Carboniferous units south and southeast of the present Winchester-Mount Sterling divide.

Mapped high-level fluvial deposits south of the divide (Irvine Formation) lie at elevations that appear too low for deposition by a postulated greater South Fork. They may have been let down from higher former deposits, but are more likely graded to the present lower Kentucky River and postdate any diversion of former South Fork headwaters into the Kentucky River basin.

However, evidence does exist for the presence of higher-level fluvial gravels that relate to the South Fork rather than to the Kentucky River. Pebbly silt, with pebbles composed of Mississippian and Pennsylvanian rocks, is exposed at Mount Sterling (Weir, 1976) and is shown as locality 2 on figure 1. The altitude of this locality is about 308 m (1010 ft). Pebbly silt is "apparently widely distributed in southern half of quadrangle as thin patches on upland flats" (Weir, 1976). Jillson (1963, p. 18, locality 16) found several pebbles derived from the Lee Formation at an altitude of 313 m (1025 ft) on and near the divide west of Winchester. This is shown as locality 1 on figure 1.

I believe that these two high-level gravel localities represent vestiges of a once extensive throughgoing South Fork system. The Mount Sterling locality may have been the locus of an abandoned segment of the system. I also believe that other gravel deposits may be present and even widespread, along parts of the postulated course shown in figure 1, and have not been recognized simply because they consist of little-travelled Lee and Newman gravel upon Lee and Newman bedrock surfaces.

## THE INTERMEDIATE VALLEY

The Teays-age Parker strath of the lower Licking River and South Fork (Luft, 1980), for nearly 85 km between Alexandria and Shawhan, is incised within a broad, readily discernible valley (figure 1). The valley surface, which is as wide as 18 km, is demarked by accordant flat-topped bedrock spurs (Luft, 1980, p.2 and fig. 4) that are 43 to 52 meters (140 to 170 feet) beneath the upland surface and 32 to 58 meters (110 to 190 feet) above the Parker strath. Because of this intermediate position, the surface, described originally by Desjardins (1934), is herewith referred to as the intermediate surface or valley. The surface--which is particularly well developed throughout Pendleton County--delimits (1) the modern Licking River valley from above its junction with the South Fork to Grants Lick, (2) most of the Teays-age Licking River valley from above its junction with the South Fork to Alexandria, and (3) nearly all of the modern and Teays-age South Fork valleys.

The surface slopes gently northward at about 0.25 m/km (figure 2), which

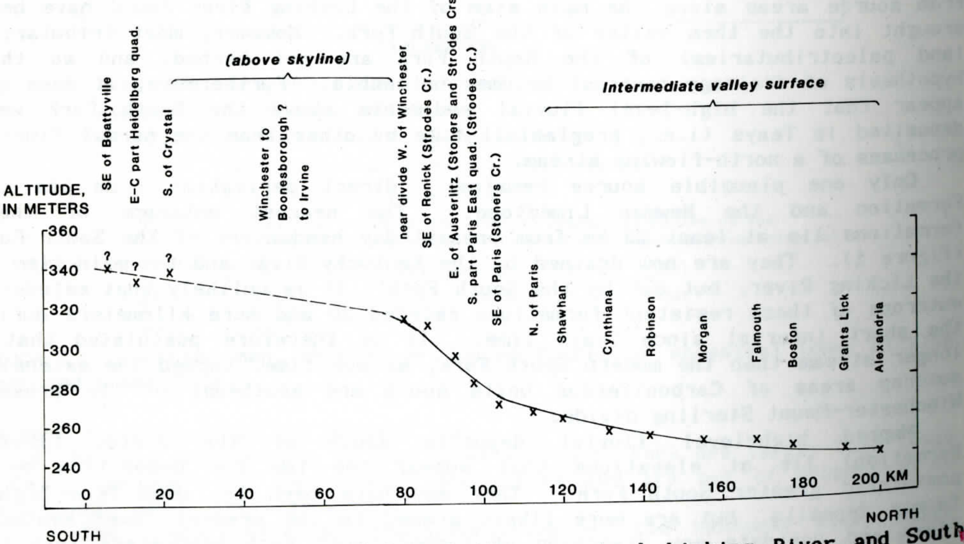


Figure 2. Longitudinal profile of the ancestral Licking River and South Fork.

approximately equals the gradient of the modern river. Such a low gradient suggests long-time formation and occupancy by a generally graded river. Throughout most of the 85-km distance, the regional northward dip of the bedrock strata, as measured from Robinson to near Alexandria, is about 1.6 m/km. From south to north, the valley surface transects increasingly younger strata, progressing from the Tanglewood Limestone Member of the Lexington Limestone (Middle Ordovician) to the lower part of the Fairview Formation (Upper Ordovician). The surface, therefore, appears to have been neither stratigraphically nor structurally controlled.

The intermediate valley, cut prior to Teays time, continues to be widened by the present river in the easily eroded shale-rich strata that form much of the valley walls. Though other processes were also at work elsewhere along the Licking River system, slope retreat was probably the dominant erosive agent along the South Fork valley, where the intermediate valley is generally of uniform width (4.6-6 km). The valley appears to have formed during a considerable amount of time beginning in the late Tertiary and continuing episodically into the present.

As originally formed the intermediate valley was only partly occupied by the ancestral river, and traces of the river's channel and bedload have been, with time, obliterated by meandering, downcutting and avulsion. For these reasons, gravel deposits are rarely found on the accordant flat-topped spurs. Such lag deposits of exotic gravel as do occur may mark the approximate position of the channel of the pre-Teays river that initiated the cutting of the intermediate valley.

The valley surface loses definition upstream, between Shawhan and Paris, and appears to be graded in that direction towards the upland surface west of Winchester (figs. 1 and 2). The lack of evidence for the presence of a postulated paleostream reach between Winchester and Beattyville in a vexing but not unsurmountable problem. From Winchester, via perhaps Boonesborough, to Irvine, the reach simply may have lain above the present land surface (figs. 1 and 2).

It is along this segment that the projection of the ancestral river crosses an area of major faulting (fig. 1). The Kentucky River fault system consists of normal faults and grabens with total vertical displacements of as much as 183 m (600 ft) (Black and Haney, 1975, p.1), or more than enough throw to smooth out all discrepancies in the longitudinal profile. Evidence for any late or post-Tertiary movement is only indirect, and the fault system appears relatively stable at the present time (Hadley and Devine, 1974). However, according to D. F. B. Black (U.S. Geological Survey, oral commun., 1984), recurrent movements along the fault system have resulted in incremental displacements, and perhaps as much as 75 percent of the total displacement could have occurred in post-Devonian time. Movement on the Cincinnati arch, a branch of which is crossed by the South Fork near Shawhan, may have been similar in magnitude and time. The major structural blocks in this area are, however, downfaulted to the south (fig. 1), or in the opposite direction from that which might have caused tectonic oversteepening of the postulated longitudinal profile upstream of Paris. Late and post-Tertiary tectonic movements in this area nevertheless may have been partly responsible for topographic irregularities in the longitudinal profile of the postulated ancestral stream. Nonetheless, it matters little whether firm evidence for or against this conjecture is ever found, because vertical movements are not absolutely necessary to explain the postulated drainage changes. In any case, faulting was responsible for the localization and trend of the Kentucky River Gorge, and rapid erosion along and near the gorge has left no topography high enough to preserve traces of a former high-level river valley.

The oversteepened reach and the apparent convexity of the longitudinal profile upstream of Paris and of the mouth of Strodes Creek (fig. 2) presently defy explanation. As indicated above, tectonic tilting was probably a negligible factor in their formation. Other possible factors that together or separately may have accounted for the anomalous profile include the interrupted or incomplete migration or removal of one or more knickpoints, the presence of resistant beds in that part of the Clays Ferry Formation and older Ordovician rocks exposed along this reach, variable discharge rates that were related to influx from tributary streams, variations in sediment size and load, variations in channel width and (or) depth and in channel sinuosity, perhaps even climatic and base-level changes, and as postulated here, an interruption in the valley's formation and its partial abandonment as a result of upriver piracy. Present knowledge precludes any reasonable assesment of these or other possible causes of the bulge in the profile.

Upstream to Beattyville, and beyond Beattyville to its ultimate beginnings, the projected extension of the intermediate valley would have had to lie upon or perhaps even above the present upland surface. All this remains completely conjectural. No river gravels at appropriate altitudes (about 335 meters (1100 feet) or more) have been reported from this southern area. Only the relative magnitude of South Fork valley meanders and the presence of giant upland meander scars near Beattyville, both to be discussed below, stand as physiographic evidence for a former throughgoing river that apparently had its sources far beyond the present Winchester divide.

The concept of the intermediate valley and the perplexing longitudinal profile shown in figure 2 pose some difficult, still unresolved problems. Very possibly the profile segments shown below and above Winchester on figure 2 may even represent two chronologically separate stages of river evolution. These problems, however, do not necessarily invalidate the premise of a longer South Fork stream draining a larger South Fork basin during late Tertiary time. It is my belief that the profile is one of a system that was only locally at grade, and that the system was disrupted by the beheading of a longer, ancestral South Fork. This event, which occurred during the planation of the intermediate valley surface and therefore, prior to Teays time, is discussed further along.

**Table 1. Meander wavelengths of the lower courses of the modern and Teays-age Licking River and South Fork.**

| River Segment | Wavelengths of Small "Stream" Meanders [1, 2] | | | Wavelengths of Large (Valley) Meanders [1, 3, 4] | | |
|---|---|---|---|---|---|---|
| | No. of measurements | Average | Range | No. of measurements | Average | Range |
| **Modern Drainage** | | | | | | |
| Lower course Licking River: near north edge Alexandria quadrangle to Falmouth | 30 | 2,890 m (9,480 ft) | 910 - 3,780 m (3,000 - 12,800 ft) | 22 | 3,750 m (12,300 ft) | 2,070 - 9,.200 m (6,800 - 30,200 ft) |
| Main stem: Falmouth to between Blue Licks Spring and Myers | 41 | 2,370 m (7,775 ft) | 790 - 5,300 m (2,600 - 17,400 ft) | | | [5] |
| South Fork: Falmouth to Cynthiana | 37 | 2,070 m (6,789 ft) | 610 - 3,380 m (2,000 - 11,200 ft) | 31 | 2,380 m (7,800 ft) | 1,100 - 5,240 m (3,600 - 17,200 ft) |
| **Teays-age Drainage** | | | | | | |
| Lower course Licking River: Pooles Creek to Falmouth | 43 | 2,950 m (9,693 ft) | 1,160 - 5,550 m (3,800 - 18,200 ft) | 23 | 4,540 m [6] (14,887 ft) | 1,765 - 8,050 m (5,800 - 26,400 ft) |
| Main stem: Falmouth to between Blue Licks Spring and Myers | 47 | 2,430 m (7,957 ft) | 1,100 - 5,360 m (3,600 - 17,600 ft) | 35 | 2,880 m [7] (9,457 ft) | 1,280 - 5,430 m (4,200 - 17,800 ft) |
| South Fork: Falmouth to northern part Cynthiana quadrangle | 30 | 1,950 m (6,400 ft) | 730 - 3,600 m (2,400 - 11,800 ft) | 12 | 3,950 m (12,967 ft) | 1,645 - 5,850 m (5,400 - 19,200 ft) |

1 Measured to nearest 200 ft on map in Luft (1980) or on geologic quadrangle maps cited therein
2 All channel bends considered to be meanders are included
3 Only meanders that approach or impinge upon valley walls (true valley meanders) are included for the modern drainage
4 Only large and (or) well-developed meanders are included for the Teays-age drainage
5 Most of segment consists of incised meanders; therefore, "stream" meanders ≅ valley meanders
6 For a part of this segment, downstream from Butler, Dury and Teller (1975, table 1) obtained a value of 4,060 m (13,200 ft)
7 The eight meanders between Falmouth and the mouth of the North Fork average 3,980 m (13,050 ft) and range from 2,930 to 5,430 m (9,600 to 17,800 ft)

## MEANDER WAVELENGTHS AND EXTENT OF DRAINAGE BASINS

Dury (1960,1965) has shown that, when plotted to logarithmic scales, a linear relation exists between meander wavelengths of rivers and the size of their drainage basins.

Meander wavelengths of the Teays-age Licking River and South Fork are shown in table 1. They are slightly to considerably larger that those of equivalent segments of the modern streams. In part, the difference in wavelength is due to the fact that some small Teays-age stream meanders were not detected and, hence, some large meanders actually may be compounds of smaller ones. All meanders of the modern streams can be readily recognized and measured on maps; those of the ancestral streams are the result of more-or-less subjective recognition and measurement. Less reliance should therefore be placed upon meanders of the ancestral streams though most large meanders appear to be in the right order of magnitude.

Average meander wavelengths of modern segments were plotted against present drainage areas on figure 3 (as black circles and squares). This figure has been derived from figure 2 of Dury (1960, p. 238) and figure 6 of Dury (1965, p. C7). Average wavelengths of the respective Teays-age segments were then projected from these points along lines drawn parallel to the boundary (defined by Dury, 1960, 1965) between valley meanders and stream meanders, and are shown by open circles and asterisks. This boundary lies parallel or nearly so to the plots of meander bands in Dury's graphs. On the basis of these constructions, the large Teays-age meanders of the Licking River and South Fork are true valley meanders. Table 1 and figure 3

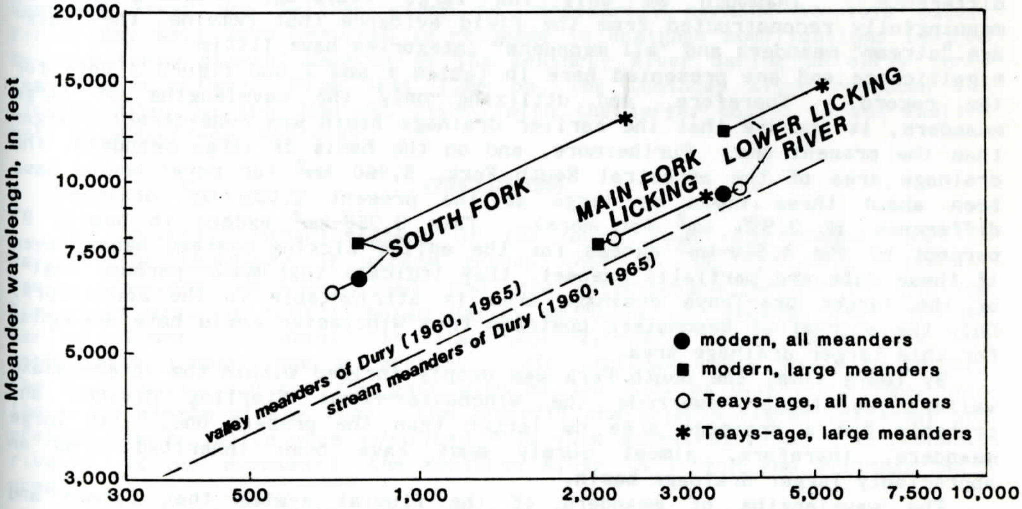


**Figure 3. Plots of wavelength against drainage area, Licking River and South Fork. Pre-Teays drainage-area projections (determined from large Teays-age meanders) may result in minimal values for those drainage areas. (Data from tables 1 and 2., To convert to meters and square kilometers, 1000 ft = 305 m. and 100 mi $^2$ = 259 km$^2$)**

indicate that large (valley) meanders of the Teays-age South Fork are considerably larger on the average, than those of the modern South Fork, and also are larger than those of the equivalent segment of the Teays-age main stem of the Licking River.

According to the projections shown on figure 3 and values derived from them for table 2, the Teays-age meanders reflected runoff from a Licking basin that was either (1) only slightly larger, on the basis of "all meanders", or else (2) considerably larger, on the basis of wavelengths of large (valley) meanders only. In the first instance, the area would have

**Table 2. Drainage areas of the modern and ancestral (late Tertiary) Licking River and South Fork.**

| River Segment | Approximate Drainage Area[1] | Approximate Drainage Area above middle of segment | Projected late-Tertiary Drainage Areas above middle of segment | |
|---|---|---|---|---|
| | | | from all Teays-age meanders | from large (valley) Teays-age meanders[2] |
| Licking River: from near north edge Alexandria quadrangle to Falmouth | 9,325 - 8,480 $km^2$ (3,600 - 3,275 $mi^2$) | 8,910 $km^2$ (3,440 $mi^2$) | 9,580 $km^2$ (3,700 $mi^2$) | 13,470 $km^2$ (5,200 $mi^2$) |
| Main stem: between Blue Licks Spring and Myers | 6,085 - 4,585 $km^2$ (2,350 - 1,770 $mi^2$) | 5,335 $km^2$ (2,060 $mi^2$) | 5,700 $km^2$ (2,200 $mi^2$) | 8,290 $km^2$ (3,200 $mi^2$) |
| South Fork: from Falmouth to Cynthiana | 2,400 - 1,610 $km^2$ (927 - 621 $mi^2$) | 2,005 $km^2$ (775 $mi^2$) | 1,810 $km^2$ (700 $mi^2$) | 5,960 $km^2$ (2,300 $mi^2$) |
| Total Licking River basin | 9,600 $km^2$ (3,707 $mi^2$) | --- | --- | >13,470 $km^2$ (>5,200 $mi^2$) |

1 From data in Speer and Gamble (1965), U.S. Geological Survey (1957), and water resources files of the U.S. Geological Survey, Louisville, Kentucky

2 These figures probably represent minimal values

been only 9,580 $km^2$ during or just prior to Teays time, compared to 8,910 $km^2$ at present (table 2), a not very significant difference. In the second instance, though, the former drainage area of more that 13,470 $km^2$ would have been at least 4,560 $km^2$ larger than the present basin--a considerable difference. Inasmuch as only the large Teays-age meanders can be meaningfully reconstructed from the field evidence that remains, the Teays-age "stream" meanders and "all meanders" categories have little significance and are presented here in tables 1 and 2 and figure 3 only for the record. Therefore, and utilizing only the wavelengths of large meanders, it appears that the earlier drainage basin was considerably larger than the present one. Furthermore, and on the basis of large meanders, the drainage area of the ancestral South Fork, 5,960 $km^2$ (or more) would have been about three times as large as the present 2,005 $km^2$ drainage, a difference of 3,955 $km^2$ (or more). This 3,955-$km^2$ excess in nearly 87 percent of the 4,560-$km^2$ excess for the entire Licking system; hence, even if these data are partially suspect, they indicate that much--perhaps most--of the larger pre-Teays drainage area is attributable to the South Fork. Only the accrual of headwaters upstream from Winchester could have accounted for this larger drainage area.

By Teays time, the South Fork was deeply incised within the intermediate valley, no longer overrode the Winchester-Mount Sterling divide, and probably had a drainage area no larger than the present one. Its large meanders, therefore, almost surely must have been inherited from an appreciably larger drainage basin.

The wavelengths of meanders of the fluvial system that formed and occupied the intermediate valley and even higher surfaces are not known. However, sinuous erosion scars that probably represent very large amplitude meanders (though they are not known to contain fluvial gravel) can be traced upon the upland surface at various altitudes throughout the Beattyville 7 1/2-minute quadrangle (R. E. Eggleton, U.S. Geological Survey, unpublished data). Particularly appropriate to the projected longitudinal profile are scars present at about 332 to 340 m (1090 to 1115 ft) altitude. Their amplitudes are at least as large as those of the Teays-age and modern meanders present considerably down-gradient to the north. Their size also suggests that the South Fork's drainage area was even larger in intermediate valley (pre-Teays) time than has been implied from Teays-age meanders.

## CAPTURE OF ANCESTRAL HEADWATERS

A glance at the map (fig. 1) will show that the lower Kentucky River and its southern extension, the Dix River, are markedly parallel to the lower Licking River and the South Fork. These parallel rivers are characterized by extensive meander belts and floodplains and are bordered by wide river terraces. Not so that stretch of the Kentucky River above the junction with the Dix River and flowing within the Kentucky River Gorge, which is oriented perpendicularly to the general trend. This is a young, deeply entrenched segment, within which meanders and terraces are poorly developed and floodplains widths are minimal. The map also shows that the Kentucky River Gorge follows major elements of the Kentucky River fault system.

I infer here that, in pre-Teays late Tertiary time, the South Fork extended into and drained the upper part of the present Kentucky River basin. It was also then eroding and transporting the Carboniferous rocks of that upper basin. Subsequently this ancestral South Fork was beheaded east of Boonesborough, and the upper basin was captured by the Kentucky-Dix River along the reach of the Kentucky River Gorge between High Bridge and Boonesborough. The South Fork thus lost its source of Carboniferous bedload.

I agree with Jillson (1963) that the Kentucky River fault system created zones of weakness in otherwise highly resistant Ordovician limestones and that these zones were followed by the Kentucky-Dix River in its act of piracy. We are also in agreement on the relatively young age (late Tertiary to perhaps very early Pleistocene, but definitely pre-Teays) of that reach of the Kentucky River that presently occupies the Kentucky River Gorge.

The ultimate reason or reasons for this postulated piracy by the Kentucky River are neither clear no necessarily germane to this paper. Piracy may well have resulted in response to drainage modifications induced by downstream glacial damming of the Kentucky River during an early (pre-Teays) glaciation. From that time on, the Kentucky River replaced the Licking River as the dominant river (with the larger basin) in the eastern half of present-day Kentucky.

## CONCLUSIONS

The underfit characteristics of the South Fork with respect to its valley, the abundance of stream gravel that is atypical of its present basin, the long, well developed, and wide intermediate earlier valley, and the large size of pre-Illinoian valley meanders all suggest that in some earlier times, presumably in the late Tertiary, the South Fork was appreciably longer than at present. Its drainage basin was also larger. The South Fork, and not the main stem of the Licking River, may have been the trunk stream of the time. The underfit characteristics of Teays-age and present meander belts were inherited from this ancestral river. It probably rivaled, if not surpassed, the Kentucky River in length and drainage-basin area.

This ancestral South Fork headed in Carboniferous rocks somewhere between Beattyville and the northwest limb of Pine Mountain (near the Kentucky-Virginia border). Piracy by the Kentucky-Dix River, along the faulted and easily eroded Kentucky River Gorge, resulted in the South Fork being beheaded. The resulting shortened South Fork, then as now heading near Winchester, incised itself through its earlier valley and formed the Teays-age Parker strath. Though now cut off from sources of Carboniferous material, it was able to transport and redistribute the vast amounts of Carboniferous gravel inherited from its pre-piracy cycle.

This act of piracy probably took place during the formation of the intermediate valley, thereby precluding its approaching graded conditions, though it could also have taken place at some later but still pre-Teays time. The system of Teays-age and younger valley meanders, and the Teays-age and younger South Fork fluvial deposits of quartz and chert, were thereby inherited from an earlier, aborted, intermediate-valley stage.

An "absolute" time frame for these events cannot be established. The Parker strath is pre-Illinoian--perhaps pre-Nebraskan or pre-Kansan--in age, having been abandoned after downriver disruption by an early Pleistocene glaciation (Luft, 1980, p. 5). It may be far older and thereby antedate the Pleistocene. The ancestral river that occupied the intermediate valley would therefore have existed in Tertiary time. However, the intermediate valley and the Parker strath also may be products of as-yet undated, pre-Nebraskan (though not necessarily pre-Pleistocene) regional glacial events, as suggested by Dury and Teller (1975).

Jillson (1963, frontispiece) delineated an ancient river course that, from Irvine to north of Winchester, is identical to the one I have proposed here. However, his postulated course continued northwestward across the present divide and was a part of the ancestral Kentucky River, whereas I suggest the course trends northward into the present valley of the South Fork. Our divergent courses may have equal merit. However, the one I propose here provided the South Fork with an ample supply of Carboniferous source material. This arguement and the others presented here--whatever their separate merits--together strongly suggest that a course such as the one depicted on figure 1 may well have existed in earlier times.

## ACKNOWLEDGEMENTS

Numerous discussions on the general subject of the pre-modern Licking River, particularly with A. B. Gibbons, and also with E. R. Cressman, K. L. Pierce, W C Swadley and J. W. Whitney, all of the U.S. Geological Survey, have helped shape my thoughts of the specific topic of the South Fork. The conclusions presented here are, however, entirely my own. Earlier versions of this paper have benefitted from reviews by A. B. Gibbons, V. L. Freeman, and M. W. Green of the U. S. Geological Survey, and J. T. Hack and Marie Morisawa.

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# PETROLOGY AND PALEOECOLOGY OF THE PACHUTA MARL (EOCENE) OF MISSISSIPPI AND ALABAMA

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## ABSTRACT

**Thin section and X-ray analyses show that the Pachuta Marl is a quartzose, fossiliferous, glauconitic, argillaceous chalk. This chalk represents a transitional facies between the western clastic Jacksonian (Late Eocene) deposits of Mississippi and the eastern calcareous deposits of Alabama and Florida. The Pachuta is generally homogeneous except for a northwestward trending increase in terrigenous material. Terrigenous components are primarily montmorillonite and quartz sand and silt, with minor amounts of kaolinite, limonite and feldspar. These components and $CaCO_3$ skeletal fragments are set in a matrix of low-Mg micrite composed primarily of calcareous nannoplankton and fragments of planktic foraminifers. Thin-section and scanning electron microscope (SEM) data suggest lithification and cementation were initiated very early in the diagenetic history of the Pachuta Marl. Early lithification started with $CaCO_3$ precipitation (as secondary overgrowths on calcareous nannoplankton and as sparry calcite cement) and ended with recrystallization (neomorphism). The lithification of lime mud and neomorphism yielded a range of fabrics (micrite, microspar, and pseudospar) which strongly resisted further diagenetic change. The cementation and recrystallization processes were inhibited by the abundance of clay minerals in the Pachuta Marl, and lithification thus resulted in a variety of substrate conditions, including the spotty occurrence of hardgrounds.**

**The faunal composition of the Pachuta Marl is dominated by molluscs, bryozoans, and echinoids. Infaunal bivalves are absent. Organisms with adaptations for soft substrates preferably inhabit those sediments containing a higher clay content. The inclusion of organisms less well adapted for life on soft substrates leads to the conclusion that the distribution of fossils in the Pachuta Marl may have been caused by a range of sediment consistencies varying from firmly consolidated to unconsolidated. The influence of clay minerals on early diagenesis in the Pachuta Marl thus plays an important role in determining the faunal composition and distribution of the Pachuta chalk substrates.**

## INTRODUCTION

Jacksonian Stage (Late Eocene) deposits of the central and eastern Gulf Coastal Plain comprise two lithofacies: a western clastic lithofacies in Mississippi and an eastern calcareous lithofacies which is typically developed in eastern Alabama and Florida (Huddlestun, 1965). The two lithofacies intergrade and intertongue across Alabama. The Jacksonian strata in southern Alabama consist of a basal clastic deposit (lower Moodys Branch Formation), a tongue of the eastern calcareous lithofacies that extends westward to the Alabama River (upper Moodys Branch Formation), a tongue of the western clastic lithofacies that extends at least as far east as the Sepulga River (North Twistwood Creek Clay, Cocoa Sand, and Pachuta Marl Members of the Yazoo Clay), and another tongue of the eastern calcareous lithofacies that extends west to the Tombigbee River (Shubuta Member and equivalent upper Crystal River Formation) Huddlestun and Toulmin, 1965) (Figure 1).

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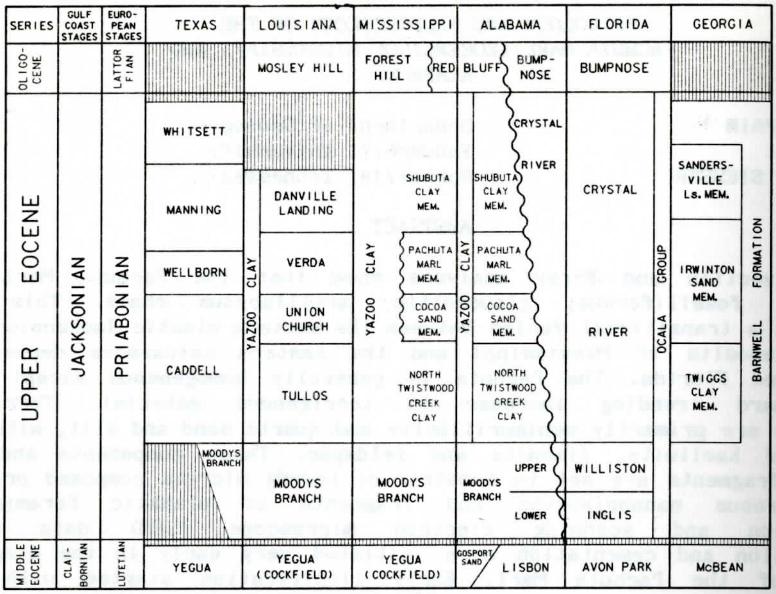


**Figure 1. Correlation chart of upper Eocene formations in the Gulf Coast Region. Modified from Huddlestun, 1965.**

The Pachuta Marl Member of the Yazoo Clay is an excellent example of the intertonguing and intergrading of the clastic and calcareous lithofacies. The purpose of this study is to investigate the petrology and aspects of the paleoecology of the Pachuta Marl Member of the Yazoo Clay. Specific intentions are to describe the vertical and lateral changes in the lithology, faunal composition, depositional environments and diagenesis of the Pachuta Marl, and to determine the relationship of fossil organisms to their environment and to one another during the stratigraphic interval represented by the Pachuta.

Sampling localities are shown on Figure 2. Stratigraphic details for each locality are given in Spain (1982). Certain field criteria were used to distinguish the Pachuta Marl from other members of the Yazoo Clay. These criteria include: 1) the presence of the echinoid *Periarchus pileussinensis*, the bivalve *Chalmys spillmani* and abundant cheilostome bryozoans in the Pachuta, and their absence in other members, 2) the abundance of $CaCO_3$ (>65%) and the relatively small amount of quartz sand (<50%) in the Pachuta compared with the Cocoa Sand Member, 3) the chalky texture of the Pachuta and the presence of abundant limonitic and glauconitic molds and casts of fossils, and 4) the use of published geologic sections to aid in differentiating similar lithologies within the Yazoo Clay.

In the area of its type locality in eastern Mississippi, the Pachuta Marl consists of 6-12 feet of light gray, irregularly indurated, quartzose, glauconitic, fossiliferous, argillaceous chalk. In Alabama, the Pachuta Marl maintains its thickness of less than 12 feet, and consists of a light greenish-gray to buff, glauconitic, argillaceous chalk to sandy, subcoquinoid limestone. The Pachuta Marl is the "Pecten-bryozoan" or "Zeuglodon" bed of earlier workers.

## PETROLOGY

### Composition

The Pachuta Marl is an argillaceous chalk (lime wackestone/mudstone) containing 65% to more than 90% $CaCO_3$. The bulk of the $CaCO_3$ is composed of

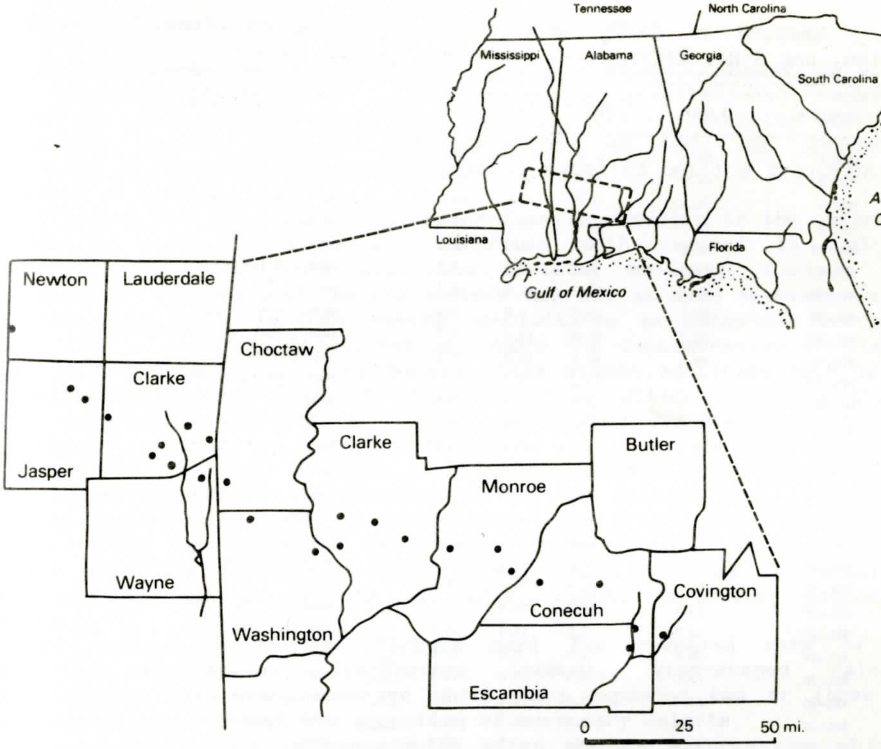


**Figure 2. Map of Eastern Gulf Coastal Region. Enlarged inset shows sample localities in Mississippi and Alabama.**

calcareous nannoplankton and planktic foraminifers. X-ray diffraction analyses show that the $CaCO_3$ is entirely low-Mg calcite (<4% $MgCO_3$). Some samples contain a significant contribution of coarser $CaCO_3$ skeletal debris including fragments of molluscs, echinoderms, bryozoa, and large benthic foraminifers. Most of these organisms were originally composed of calcite. Aragonite or high-Mg calcite were not detected in any of the samples analyzed during this study.

The acid-insoluble residues in the Pachuta contain silt and very fine sand size quartz with minor amounts of other minerals such as plagioclase feldspar and biotite, glauconite, limonite, pyrite, and clay minerals. X-ray diffraction analyses show that montmorillonite is the dominant clay mineral in all of the deposits of the Pachuta; kaolinite and other clay minerals such as chlorite and illite, where present, are markedly subordinate. As used in this paper, the term "clay" refers essentially to the mineral montmorillonite. Other insolubles are generally of only minor importance, and those contributing to the Pachuta (quartz, glauconite) are characteristically only significant in nearshore areas of chalk deposition.

Results of $CaCO_3$, X-ray, and insoluble-residue analyses of the Pachuta Marl are given in Table 1.

## Diagenesis

Calcareous nannoplankton and planktic foraminifers which make up the bulk of the Pachuta Marl are composed of low-Mg calcite. This is the most stable polymorph of $CaCO_3$ under near surface temperature and pressure conditions (Scholle, 1977). Many coarser skeletal fragments which contribute to the Pachuta Marl (bryozoans, oysters, and pectens) are

Table 1. Results of $CaCO_3$ Weight-loss Determination, Insoluble Residue Examination, and X-Ray Diffraction Analyses, Pachuta Marl.

| Sample # | % $CaCO_3$ | % Insoluble Residue | | | | X-Ray Analyses |
|---|---|---|---|---|---|---|
| | | Qtz. | Glauc. | Clay | Other | Clay Minerals Present |
| MJA-1 | 65.5 | 10.0 | tr | 25.0 | ___ | mont., chlor. (tr) |
| MJA-3 | 65.0 | 3.0 | tr | 32.0 | ___ | mont., chlor. (tr) |
| MCL-5 | 76.8 | 5.0 | 5.0 | 13.2 | ___ | mont., kaol. (tr) |
| MCL-6 | 16.0 | 55.0 | tr | 28.3 | ___ | No Data |
| MCL-7B | 74.6 | 3.0 | 3.0 | 24.1 | ___ | mont. |
| MCL-8 | 74.5 | 8.0 | 9.0 | 8.5 | ___ | mont., kaol (tr) |
| MCL-9A | 67.8 | 10.0 | 15.0 | 9.2 | ___ | mont. |
| MCL-9B | 82.2 | 5.0 | 5.0 | 8.0 | ___ | mont., kaol (tr) |
| MCL-9C | 58.7 | 20.0 | 15.0 | 8.0 | ___ | mont., kaol (tr), illite (tr) |
| MCL-9E | 59.5 | 10.0 | 15.0 | 15.0 | ___ | mont., chlor (tr) |
| MCL-9F | 59.7 | 17.0 | 20.0 | 3.0 | ___ | mont., illite (tr) |
| MCL-9G | 53.1 | 18.0 | 21.0 | 7.0 | ___ | mont. |
| MCL-9H | 71.1 | 13.0 | 12.0 | 3.9 | ___ | mont. |
| MCL-11 | 59.4 | 7.0 | 10.0 | 23.6 | ___ | mont., kaol (tr) |
| MWA-1A | 53.9 | 30.0 | 10.0 | 5.4 | ___ | mont. |
| MWA-1B | 73.1 | 8.0 | 5.0 | 8.0 | ___ | mont., chlor (tr) |
| ACH-1 | 67.0 | 19.0 | 2.0 | 12.0 | ___ | No Data |
| AWA-1 | 53.2 | 10.0 | 10.0 | 26.8 | ___ | mont. |
| AWA-2 | 86.1 | 5.0 | 3.0 | 4.8 | ___ | mont., kaol (tr) |
| ACL-1A | 59.1 | 21.0 | 10.0 | 9.0 | ___ | mont. |
| ACL-1B | 63.1 | 20.0 | 10.0 | 7.0 | ___ | mont., illite (tr) |
| ACL-1C | 70.7 | 5.0 | 15.0 | 10.0 | ___ | mont., illite (tr) |
| ACL-2A | 80.4 | 3.0 | 5.0 | 12.0 | ___ | mont. |
| ACL-2B | 74.0 | 5.0 | 8.0 | 10.7 | ___ | mont., illite (tr) |
| ACL-2C | 83.6 | 5.0 | 3.0 | 8.3 | ___ | mont., illite? (tr) |
| ACL-3A | 76.6 | 8.0 | 6.0 | 11.0 | ___ | mont., chlor (tr) |
| ACL-3B | 75.6 | 10.0 | 5.0 | 10.0 | ___ | mont. |
| ACL-3C | 78.2 | 5.0 | 3.0 | 14.0 | ___ | mont. |
| ACL-5 | 78.2 | 4.0 | 3.0 | 14.8 | ___ | mont., kaol (tr) |
| AMO-2A | 69.5 | 5.0 | 15.0 | 10.5 | ___ | mont., illite (tr) |
| AMO-2B | 81.9 | 3.0 | 5.0 | 12.1 | ___ | mont. |
| AMO-3A | 68.4 | 3.0 | 17.0 | 12.0 | ___ | mont. |
| AMO-3B | 61.6 | 15.0 | 17.0 | 7.5 | ___ | mont., kaol (tr) |
| AMO-3C | 62.4 | 8.0 | 13.0 | 16.3 | ___ | mont., kaol (tr) |
| ACO-1A | 51.1 | 25.0 | 3.0 | 21.0 | ___ | mont. |
| ACO-1B | 59.6 | 15.0 | 1.0 | 25.0 | ___ | mont., kaol (tr) |
| ACO-1C | 60.3 | 16.0 | 1.0 | 23.5 | ___ | mont. |
| ACO-2 | 89.3 | ___ | ___ | 10.6 | ___ | mont. |
| AES-1A | 87.6 | tr | 1.0 | 12.4 | ___ | mont. |
| ACOV-1A | 86.9 | tr | 1.0 | 13.1 | ___ | mont. |
| ACOV-1B | 92.2 | tr | tr | 8.0 | ___ | kaol (tr) illite (tr) |

normally completely or partially composed of low-Mg calcite (Bathurst, 1971). Because of this chemical stability, the Pachuta Marl differs greatly from most shallow-water limestones which originally contain large amounts of unstable aragonite and high-Mg calcite.

The diagenetic history of most chalks includes two stages: (1) an early diagenetic stage or lithification in which the sediment is transformed into a stable substrate by cementation at a time when the sediment is still in direct connection with the marine environment; and (2) a late (burial) diagenetic stage in which burial or post-burial alteration results from the addition of overburden by sediment accumulation. (Scholle, 1977).

Early diagenesis takes place in the shallow burial realm (1-200 meters sub-bottom) and is characterized by the effects of gravitational compaction (establishment of firm grain contacts), the dissolution of fossils and initiation of overgrowths (Schlanger and Douglas, 1974). Late diagenesis begins after subsequent burial (600-1000 meters) allows compaction, pressure solution, and recrystallization to consolidate the sediment (Wise, 1977).

## Early Diagenesis

For most chalks, early diagenesis consists of an early dewatering stage. The dominant mechanism in this realm is gravitational compaction; cementation is a subordinate process (Scholle, 1977). With the addition of overburden through continued sedimentation, as well as through the activities of burrowing organisms, water is expelled until a grain-supported framework is formed.

Following shallow burial, calcareous fossils dissolve to the extent that surrounding pore waters become saturated with respect to $CaCO_3$ and equilibrium is established with the skeletal material. Certain robust nannoplankton species such as discoasters may act as seed crystals on which the dissolved $CaCO_3$ in pore waters precipitates as secondary overgrowths (Wise, 1977). As this precipitation lowers the concentration of dissolved $CaCO_3$ in the interstitial pore waters, less stable particles such as small nannoplankton elements begin to dissolve in an attempt to bring the pore waters back up to equilibrium. Thus, the primary cementation process acts through the solution of less stable, very small calcite crystals such as thos that make up small elements of individual nannoplankton species and the walls of foraminifers, and reprecipitation for calcite upon large cystals such as those which make up discoasters and other robust nannoplankton species. The very abundant micron-sized elements can supply much of the calcite available for interstitial cement and infillings of foraminifers tests as well as overgrowths on large, robust calareous nannoplankton species.

SEM examination of the Pachuta Marl has revealed many of these diagenetic effects. Discoasters (robust, star-shaped calcareous nannoplankton) have been observed which have expanded two or three times their normal size through the accretion of secondary calcite.

The secondary calcite overgrowths often obscure bifurcations which are normally visible at ray tips, and euhedral calcite crystal faces are typically developed along the rays of the discoasters. In the Pachuta Marl, adjacent shields of coccoliths are often fused together by overgrowths of secondary calcite. Other, less stable coccoliths have undergone extensive dissolution, and their shields are frequently separated and their central elements are missing.

The extent of this selective dissolution depends on the rate of sediment supply versus the rate of sediment removal. Therefore, areas which have low rates of carbonate supply or accumulation commonly have the most extensive dissolution of the lower stability faunal elements (Scholle, 1977). Depositional hiatuses expose surface sediment to prolonged burrowing activity and dissolution by bottom waters. Other diagenetic processes are often associated with this selective seafloor dissolution, including carbonate cementation and replacement of carbonate by glauconite or phosphate. Such zones of early seafloor induration are commonly referred to as "hardgrounds" (Bathurst, 1971: Milliman, 1977).

Samples from St. Stephens Quarry in Washington County, Alabama, certainly indicate the spotty occurrence of hardgrounds in the Pachuta Marl. Phosphate nodules are abundant. Glauconite has frequently replaced carbonate grains. Borings are present which transect grain boundaries. Alcyonarians and encrusting bryozoans are abundant, also indicating an early induration of the Pachuta Marl.

Examination of thin-sections also provides evidence which suggests an early lithification of the Pachuta Marl. The fine-grained textures of the Pachuta contain abundant, tiny, sediment-filled, undeformed shells of fossils. These shells have preserved their original shape in the lithified matrix, and voids often contain sparry calcite cement and neomorphic calcite. The shells are not compressed or flattened, so they must have been supported by a rigid, grain-supporting framework before the pressure of additional overburden occurred.

## Burial Diagenesis

Features which indicate lithification in a later diagenetic stage (burial diagenesis) are absent in the Pachuta Marl. There are two main mechanisms which operate during burial diagenesis: (1) mechanical compaction, and (2) chemical compaction (Scholle, 1977). Although considerable breakage of foraminifer and ostracode tests has been observed in the Pachuta Marl, mechanical compaction appears to be only a minor diagenetic factor. The broken tests of microfossils can be explained through fragmentation by plankton feeders in the open water column or by the burrowing activities of organisms at the sediment/water interface.

Solution seams (concentrated zones of irregular, wispy, marly, and clayey sediment), stylolites, grain embayments, and other forms of evidence of chemical compaction are absent in the Pachuta Marl.

A fairly rigid framework is believed to have formed during the compactional dewatering stage of early diagenesis so that post-depositional, burial compactional effects are not observed in the Pachuta Marl. Early lithification started with $CaCO_3$ precipitation (as secondary overgrowths and sparry cement) and probably was finished by recrystallization (neomorphism) in the sediment.

## Neomorphism

The cementation and recrystallization process may be greatly inhibited by the presence of non-carbonate impurities such as clay minerals (Scholle, 1974). Where clay contents are higher, secondary overgrowths of $CaCO_3$ on larger calcareous nannoplankton may be significantly reduced because of the insulating effects of clay particles on individual nannofossils (Wise, 1977). This insulating clay layer tends to inhibit the diffusion of $CaCO_3$ necessary for the dissolution-reprecipitation reactions to occur. Consequently, besides cementation, recrystallization has to be considered also as one of the main processes during early lithification of the Pachuta Marl.

Aggrading neomorphism is defined as the process whereby finer crystal mosaics are replaced by coarser crystal mosaics of the same mineral or its polymorph, *in situ* (Folk, 1965). Many data support recrystallization by aggrading neomorphism rather than the secondary precipitation of sparry calcite as a causative agent producing the coarse crystal mosaics observed in the Pachuta Marl. These include the irregular distribution of crystal size, including micrite (0.5-4μm), microspar (5-50μm), and pseudospar (50-100μm); abrupt contacts between micrite and neomorphic spar; curved to wavy intercrystalline boundaries in the coarse, crystalline calcite mosaic; absence of enfacial junctions in a calcite mosaic; and syntaxial overgrowths of allochems in mudstones and wackestones (if the allochems are mud-supported, then the sediment was deposited in a single act and the syntaxial rim must be an *in situ* replacement) (Bathurst, 1971). Thus the possibility that the majority of coarse crystalline calcite in the Pachuta Marl is a cavity-filling sparry calcite cement or a rim-cemented deposit of detrital crystals is ruled out. The lithification of lime mud and the growth of neomorphic spar has yielded the range of low-Mg calcite fabrics (microspar, pseudospar) which strongly resist any further diagenetic change in the Pachuta.

Because the action of neomorphism is dependent on unstable minerals (Bathurst, 1971), the development of neomorphic fabrics is directly related to the development of lithification in the Pachuta Marl. The longer lithification is delayed, the more time there is for neomorphism to proceed (Bathrust 1971). Thus, the degree of neomorphism may be directly affected by the presence of clay minerals in the Pachuta which tend to inhibit the dissolution of calcite from small, less stable microfossil elements.

## PALEOCOLOGY

### Introduction

Presence-absence data, rather than abundance data, are utilized in the analysis of the Pachuta Marl fauna. Most specimens were identified to the species level, and this level is used to analyze paleoecological relationships in the Pachuta Marl. Most samples contain some or all of the species most commonly associated with the Pachuta Marl, namely those with reference numbers 1, 3, 5 and 14 listed in Table 2 and Figure 3, whereas many samples contain a few to several of the remaining 25 species. Almost all localities contained a diverse bryozoan fauna, although they are not considered in the Pachuta benthic macroinvertebrate fauna.

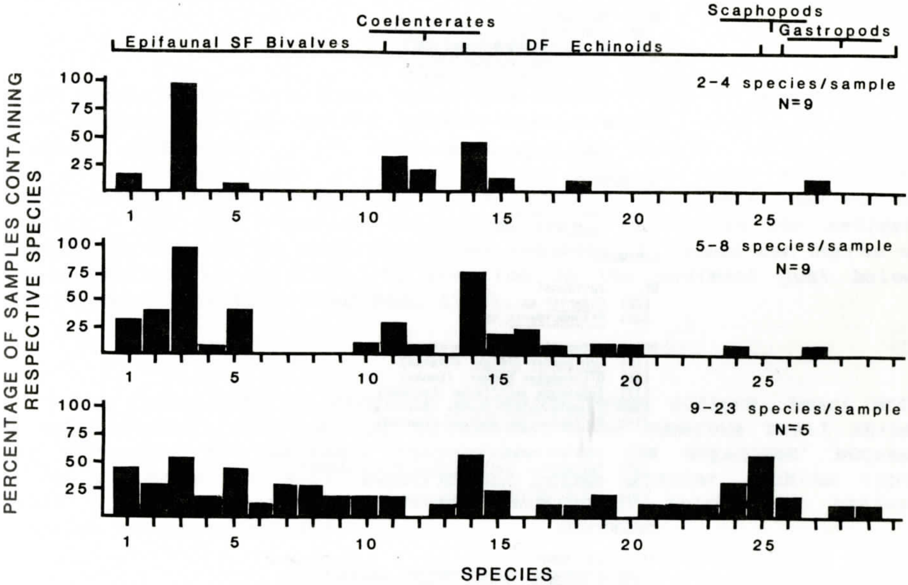


**Figure 3. Relationship between taxonomic composition and life habit composition with increase in diversity, Pachuta Marl. N is number of samples studied for each level of diversity. Key to numbers identifying species is in Table 1. Adapted from Bottjer (1981).**

The benthic macroinvertebrates of the Pachuta Marl are listed in Table 2. The trophic categorization follows terminology defined by Walker and Bambach (1974). Other paleoecological terms (organism-substrate relations) are defined by West (1977). The life habits of the bivalves, gastropods, and echinoids were evaluated from the study of data compiled from the life habits of their recent descendents and by comparison with studies of similar fossil taxa (e.g., Barnes, 1974; Bottjer, 1981; Cox and others, 1969; Kauffman, 1969; Schafer, 1972; Stanley, 1970). Interpretations of the fossils are based primarily on adult forms. The following sources were utilized in the identification of the Pachuta Marl benthic macroinvertebrate fauna: Cox and others, (1969), Dockery (1980), Gardner (1939), Harry and Dockery (1983), Hickson (1938), Howe (1937), Moore and others (1952), Thurmond and Jones (1981), Toulmin (1977), and Vaughn (1900).

### Life Habits/Functional Morphology

Many of the benthic epifaunal organisms in the Pachuta Marl show adaptations for living on soft, unstable substrates. Adaptations of skeletal

**Table 2. Life Habit Categorization of Pachuta Marl Benthic Macroinvertebrates[1].**

BIVALVES

SF, epifaunal reclining
- (1) Pycnodonta trigonalis (Conrad)
- (2) Gryphaeostrea vomer plicatella (Morton)
- (7) Capulus americanus Conrad

SF, epifaunal cemented
- (4) Pycnodonta vicksburgensis (Conrad)
- (5) Ostrea podagrina Dall
- (6) Ostrea falco Dall

SF, epifaunal bysally-attached
- (3) Chlamys (Aequipecten) spillmani (Gabb)

SF, epifaunal unattached/swimming
- (3) Chlamys (Aequipecten) spillmani (Gabb)
- (9) Chlamys spillmani clinchfieldensis Harris
- (10) Chlamys danvillenis Weisbord in Tucker-Rowland

GASTROPODS

DF, epifaunal
- (8) Crepidula sp.
- (25) Cirostrema ranellinum (Dall)
- (26) Cirostrema nassulum creolum Palmer
- (27) ?Tornatellaea lata (Conrad)
- (28) Solariella sp.

Carnivores, epifaunal

- (29) ?Conus sauridens Conrad

SCAPHOPODS

DF, semi-infaunal
- (24) Fustiaria danai (Meyer)

ECHINOIDS

DF, epifaunal
- (22) Cidaris? sp.
- (23) Prionocidaris sp.

DF, infaunal
- (15) Schizaster armiger (Clarke)
- (16) Ditremaster beckeri (Cooke)
- (17) Eupatagus antillarum (Cotteau)
- (18) Euhodia patelliformis (Bouve)
- (19) Macropneustes mortoni (Conrad)

DF, semi-infaunal
- (14) Periarchus lyelli pileussinensis (Ravenel)
- (20) Periarchus protuberans Twitchell
- (21) Weisbordella cubae (Weisbord)

COELENTERATES

SF, epifaunal
- (12) Endopachys shaleri Vaughn
- (11) Flabellum cuneiforme spp.
- (13) Eogorgia sullivani Hickson

[1]SF is suspension-feeding, DF is deposit-feeding, number in brackets before each taxon is reference number used to key-in species in Figure 13. Trophic categorization after Walker and Bambach (1974). Data obtained from Barnes, 1974; Cox et al, 1969; Harry and Dockery, 1983; Kauffman, 1969; Schafer, 1972; Stanley, 1970; and Toulmin, 1977.

structures for living on such substrates are outlined by Thayer (1975). These include: (1) reduction in body density during growth, (2) sinking into the underlying sediment until the weight of the organism equals that of the displaced sediment, (3) increasing the bearing surface by flattening and (4) maintaining small size to keep at a minimum the force per unit area applied by the organism.

## Bivalves

Interpretation of morphologic adaptations of the bivalve *Chlamys (Aequipecten) spillmani* is based on juvenile forms as well as adult forms, because the species has a characteristic swimming habit in its early ontogenetic stages. As an adult, the small, lightweight shell and active swimming of *Chlamys* sp. is adaptive to the preferred soft-mud habitat, enabling these pectens to "float" on a soft substrate when not dwelling in fissures or byssally attached to a larger, harder substrate (Stanley, 1970).

*Chlamys spillmani clinchfieldensis* and *Chlamys danvillensis* as well as some *Chlamys (Aequipecten) spillmani* may have possessed the best adaptation for life on soft substrates, because if they began to sink into the substrate, they were probably able to swim away.

Many specimens of *Pycnodonta trigonalis* exhibit deeply convex lower valves, topped by flat upper valves. This is probably an "iceberg" adaptation to soft substrates, because not all sample localities contained specimens with this adaptation. *Gryphaeostrea vomer plicatella* possesses a thick curved, deep lower valve which probably floated in or on the soft, unstable ooze. The typical incurved beak of the gryph-shaped lower valve facilitated the distribution of the load exerted by the organism's weight, and prevented sinking into the substrate. *Pycnodonta vicksburgensis* and *Ostrea podagrina* both have somewhat short spines located in the furrows of their radial ridges, which may have helped to stabilize these species in the Pachuta chalk substrate.

## Gastropods

The gastropods *Cirostrema ranellinum* (Dall) and *C. nassulum creolum* Palmer possess a high spired axially-ribbed shell, which is probably a "showshoe" adaptation to the organism's biotope (Thayer, 1975). The tusk-shape and burrowing habit of the scaphopod *Fustiaria danai* are adaptive to a soft bottom habitat; its small size reduces the force exerted on the substrate. If the organism begins to sink too far in the sediment or begins to be covered by newly deposited sediment, *F. danai* can burrow upward like a bivalve to maintain its position in the sediment just below the sediment/water interface (Kauffman, 1969).

## Echinoids

In the Pachuta Marl, irregular echinoids (sand dollars, heart urchins) are more abundant than regular echinoids. The numerous small spines on these echinoids are probably responsible for the organisms' success as shallow burrowers in soft substrates. Most regular urchins typically inhabit solid or rocky substrates where their thick long spines aid locomotion in these habitats.

## ORGANISM-SUBSTRATE RELATIONS

The amounts of clay-size matrix, quartz sand and silt, and invertebrate skeletal $CaCO_3$ grains were estimated through thin-section and binocular microscope examinations and were compared to the number of species for most of the samples. The results of this comparison are shown in Figures 4 and 5. In Figure 4, percentages of clay-size matrix and sand/silt- size grains (including quartz, glauconite, and skeletal $CaCO_3$) were plotted against sample diversity for 23 samples. Diversity is lowest in those samples containing a relatively high percentage of clay-size grains, and highest in those samples containing a high sand/silt-size grain content. This suggests that, to some degree, diversity is related to the coarseness of the substrate. Samples MCL-10 and MCL-6 are from the lower Pachuta Marl in Clarke County, Mississippi. In this area, the lower Pachuta is considered to be the equivalent of the Cocoa Sand, which explains the unusually high sand content.

Diversity may be also related to the amount of clay minerals present in the samples. In Figure 5, the percentage of clay minerals was compared to the number of species in the same samples. The graph shows that those samples containing a relatively high amount of clay minerals are those which exhibit low sample diversity. These samples primarily contain fossils which are especially adapted for life on soft, generally unconsolidated substrates (Figure 3).

The increase in life habit variety, with increase in sample diversity

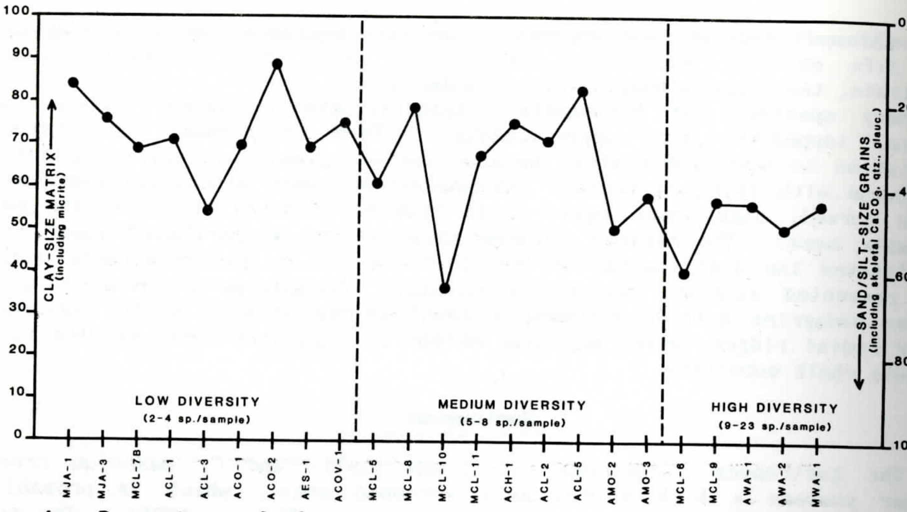


**Figure 4. Percentage of clay-size matrix and sand/silt-size grains compared to sample diversity, Pachuta Marl.**

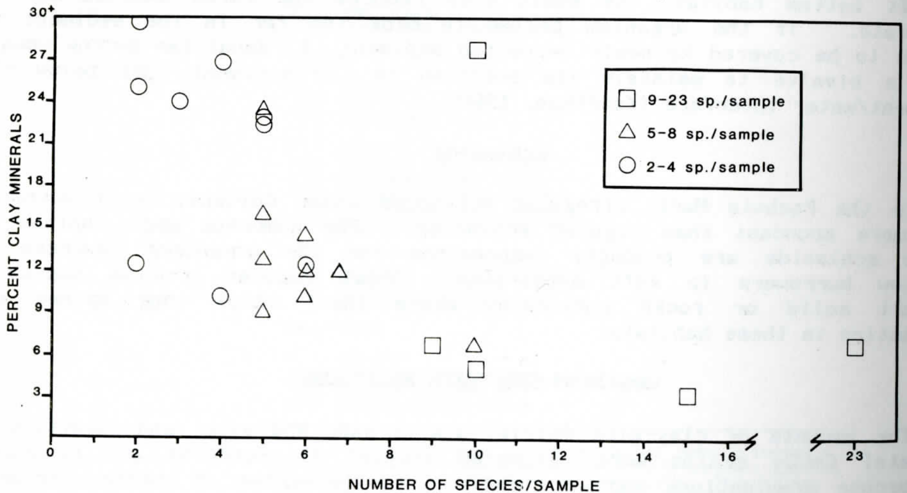


**Figure 5. Percentage of clay minerals compared to the number of species for samples in Figure 14.**

seen in Figure 3, can also be explained by the aspect of variable substrate consistencies. This increase in life habit variety can be outlined as follows.

1. Low diversity (2-4 species/sample): shallow burrowing echinoids, small solitary corals, and epifaunal suspension feeding bivalves with adaptations for life on soft substrates. *Ostrea podagrina* and *Tornatella lata* are also present.

2. Medium diversity (5-8 species/sample): The commonly occurring echinoids and soft-bodied infaunal deposit feeders characteristic of the low diversity samples are still present, as well as the epifaunal suspension feeding bivalves. These are joined by less commonly occurring bivalves also adapted to life on soft substrates (*Gryphaeostrea vomer plicatella, Pycnondonta vicksburgensis*).

*Ostrea falco* is also present, which probably attached to the reclining bivalves or to a stable substrate such as shell fragments. Infaunal deposit-feeding echinoids are joined by *Ditremaster beckeri*, *Eupatagus antillarum*, and *Macropneustes mortoni*. Gastropods and scaphopods are rare. Infaunal deposit-feeding bivalves are absent.

3. High diversity (9-23 species/sample): The characteristic epifaunal suspension-feeding bivalve fauna with adaptations for soft substrates as well as the commonly occurring coelenterate and echinoid fauna are still present. The diversity increased with the addition of organisms better adapted for life on consolidated substrates, particularly the alcyonarian *Eogorgia sullivani* and abundant encrusting organisms. The bivalve fauna was completed with the addition of cemented epifaunal suspension feeders and small swimming pectinaceans. The swimming habit of *Chlamys spillmani* may have been an escape mechanism to avoid predation in a highly diverse fauna. Epifaunal echinoids are present. Gastropod diversity also increased, and included the carnivorous species *?Conus sauridens*. Infaunal deposit feeding bivalves are absent.

The progressive inclusion of organisms less well adapted for life on soft substrates leads to the conclusion that the distribution of fossils in the Pachuta Marl may have been caused by a range of substrate consistencies varying from firmly consolidated to unconsolidated.

## CONCLUSIONS

The Pachuta Marl (Late Eocene) is a quartzose, glauconitic, argillaceous chalk (lime wackestone/mudstone). The Pachuta Marl is dominated by molluscs, bryozoans, and echinoids, namely *Pycnodonta trigonalis*, *Chlamys spillmani*, *Pycnodonta vicksburgensis*, *Gryphaeostrea vomer plicatella*, *Ostrea podagrina*, *Cirsotrema nassulum creolum*, *Periarchus lyelli pileussinensis*, and *Schizaster armiger*. Vertebrate fragments are also common in the Pachuta Marl. Many of the benthic macroinvertebrate fauna of the Pachuta exhibit morphologic adaptations for living on soft, unstable substrates.

Calcareous nannoplankton and planktic foraminifers, which comprise the bulk of the lime wackestones and packstones of the Pachuta Marl, are composed of low-Mg calcite. Neither aragonite nor high-Mg calcite were detected in any of the samples analyzed during this study. Because of this primary chemical stability, freshwater exposure at near-surface conditions have had little effect on the diagenesis of the Pachuta Marl. Lithification and cementation were initiated very early in the diagenetic history of the Pachuta. Early diagenesis began with a gravitational, compactional dewatering stage in which firm grain contacts were established. Dissolution of less stable calcite crystals (Small calcareous nannoplankton elements and walls of foraminifers) and reprecipitation of calcite as secondary overgrowths on larger nannoplankton species created the interstitial calcite cement and infillings of foraminifer tests in the Pachuta Marl. A fairly rigid framework was formed during this compositional, dewatering stage.

Early lithification in the Pachuta thus started with $CaCO_3$ precipitation (as secondary overgrowths and sparry calcite cement) and ended with recrystallization (neomorphism) of lime mud. The lithification of lime mud and the growth of neomorphic spar yielded a range of fabrics (microspar, pseudospar) which strongly resisted any further diagenetic change.

The cementation and recrystallization process was greatly inhibited by the abundance of clay minerals in the Pachuta Marl. Among other factors, relatively large amounts of clay in a sediment tend to reduce the precipitation of secondary calcite overgrowths on calcareous nannoplankton (Wise, 1977), because of the insulating effects of clay particles on individual nannoplankton.

The variability of substrate consistencies may also be related to the amount of clay minerals present in the Pachuta Marl. Those samples of the Pachuta which contain a high clay content are typically inhabited by fewer benthic macroinvertebrate species, especially those fossils exhibiting morphologic adaptations to soft substrates. Samples containing a low clay content are inhabited by a greater number of fossil species, including a number of specimens which are characteristically well adapted for life on firm substrates, leading to the conclusion that the distribution of fossils in the Pachuta Marl may have been caused by a range of substrate consistencies. The influence of clay minerals on early diagenesis plays an indirect role in determining the faunal composition and distribution of the Pachuta chalk substrates.

## ACKNOWLEDGEMENTS

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## TERRACE STRATIGRAPHY ALONG THE LOWER RED RIVER, LOUISIANA: DISCUSSION AND REPLY

### DISCUSSION

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In a recent paper, Alford and others (1985) question the acceptance of a Sangamonian age for the Gulf Coast Prairie Formation and on the basis of radiocarbon dates from a Fiskian "pre-Prairie" river terrace prefer a mid-Wisconsinan age instead. It was Fisk (1938a) who designated as Prairie a low depositional stream terrace in central-southwestern Louisiana, correlating it with a southeastern Louisiana coastwise terrace surface, formed from overlapping floodplains of adjacent streams near the seashore. Without tackling the very involved age issue of the "Prairie terrace" along the Mississippi and major tributaries, as well as its correlatability with Fisk's coastwise Prairie, I had attempted to show the Sangamonian age of the alluvial unit that directly underlies the extensive and well-definable, youngest Pleistocene coastwise surface along the northern Gulf Coast.

East of the Mississippi River, this surface (also called Pamlico in western Florida and Beaumont in Texas) is bounded by the late Pliocene-early Pleistocene (?) Citronelle upland and small remnants of older Pleistocene alluvial units; seaward by the late Pleistocene Gulfport-Ingleside barrier plain and Holocene coastal units. Inland from the present shoreline, the Prairie sequence interfingers with and is underlain by the Biloxi Formation that displays a broad range of open marine-to-highly brackish biotopes, and directly underlies the Gulfport coastal barrier complex (Figure 1).

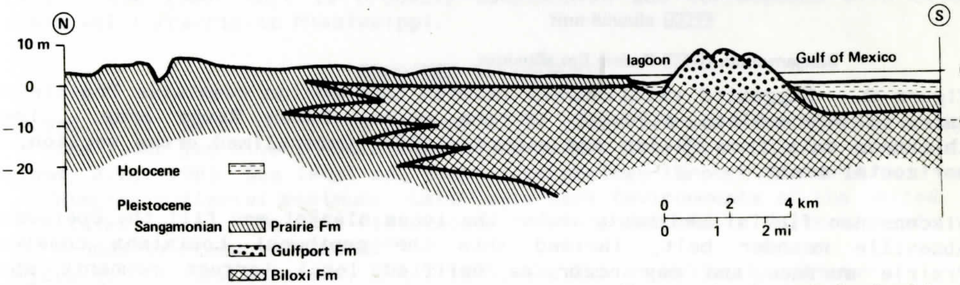


**Figure 1. Generalized cross section between Mississippi and Apalachicola Rivers, northeastern Gulf Coast, indicating interrelationship between Sangamonian coastwise Prairie, Gulfport barrier, and estuarine-to-open nearshore marine Biloxi Formation (Otvos, 1972a, 1985). With falling late Sangamonian-early Wisconsinan sea level, time-transgressive deposition of Prairie alluvium continued off the present mainland shore.**

Deposition of the three interrelated units took place during the Sangamonian high sea level stage. (+6 - +7m). At −38 to −42m, the highest assumed Wisconsinan interstadial sea level (Bloom, 1983) remained well below the current one and could not have been associated with these coastal and marine formations. The Biloxi, traced landward from inner shelf deposits beneath the present shelf to an irregular estuarine feather edge slightly above current sea level well inland under the present coastal plain, represents the transgressive-regressive marine cycle during the last interglacial (Otvos, 1972a, 1975, 1985). This Formation has been encountered in hundreds of coreholes in the shallow subsurface, practically along the entire length of the coastal plain.

Presistent buried soil horizons or their absence in loess that blankets a given Pleistocene alluvial unit may indeed be helpful in defining the minimum age of the unit, provided subsequent erosion did not prevent preservation of the enclosed soil horizon. However, the lack of soil zones within a well-developed loess sequence even in a low-relief coastal plain setting, as at Baton Rouge and Lafayette, Louisiana does not necessarily call for designation of the underlying Wisconsinan alluvial unit as the (coastwise) Prairie Formation. Instead, such units may represent younger Pleistocene deposits. I suggest that at the valley scarp (Figure 2) Wisconsinan-age river deposits, not the Prairie, underlie the buried soil-free, 4.2-m "Peorian" loess (Sites 6 and 9; Miller and others, 1982).

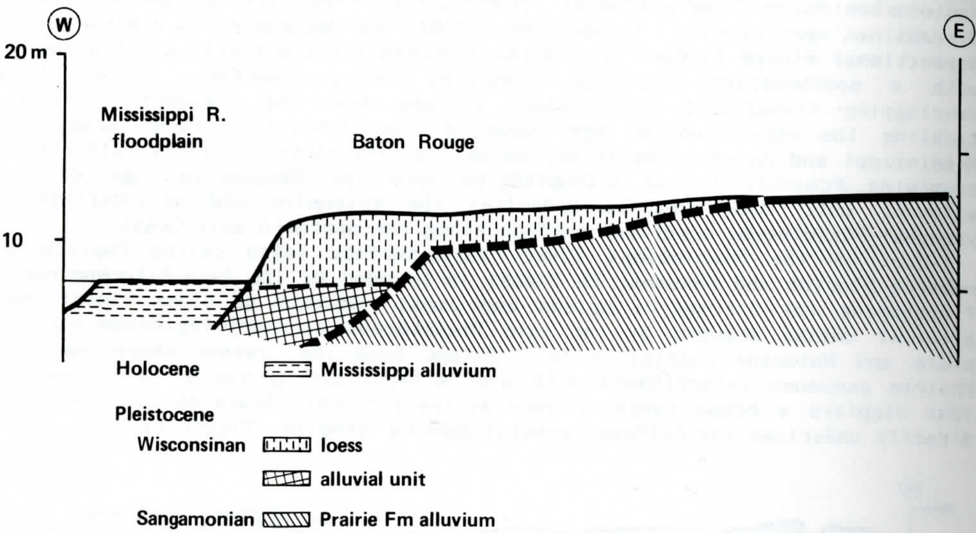


**Figure 2. Suggested relationship between Sangamonian coastwise Prairie and inferred Wisconsinan alluvial terrace units, Baton Rouge area. Loess thickness data from Miller and others (1982). Generalized cross section, no horizontal scale.**

Wisconsinan fluvial sediments under the loess blanket may fill the Opelousas-Abbeville meander belt, incised into the southwest Louisiana coastwise Prairie surface and may occur as uplifted local terrace remnants above present floodplain level in Mississippi River bluffs upstream from Baton Rouge.

## REPLY TO DISCUSSION


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## AGE OF THE PRAIRIE TERRACE

Although it has become fashionable to consider the Prairie a Sangamonian terrace (Gagliano and Thom, 1967; Hoyt and others, 1968; Saucier, 1968; Otvos, 1972b; Delcourt and Delcourt, 1977) the validity of this assignment depends upon location. Our dates from Waddel Bluff provide strong evidence that the Prairie Terrace fill in the lower Red River Valley is late Altonian to Farmdalian in age (Alford and others, 1985). In other areas fills that have been mapped as Prairie have produced only "dead" dates and are presumably Sangamonian. From a time stratigraphic point of view it is not very satisfactory to have a single terrace representing all the cut and fill cycles that occurred between 25,000 B.P. and 125,000 B.P.

There are several solutions to this dilemma. Certainly interested parties could agree that the name Prairie be restricted to fills of the Sangamonian Stage. Although this step might be widely accepted it runs contrary to geologic protocol. In the type area, Fisk (1938b) considered the Prairie to be mid-Wisconsinan in age and since the $^{14}C$ record indicates that this assignment was correct it would be more orderly to reserve the Prairie appellation for those fills that occupy the mid-Wisconsinan time slot.

Protocol aside, the problem with this scheme is the substantial body of literature (eg. Gagliano and Thom, 1967; Saucier, 1968; Saucier and Fleetwood, 1970) that has placed the Deweyville into the mid-Wisconsinan slot. With such a history of confusion it might be best to drop the Fiskian nomenclature from time stratigraphic studies and to simply refer to given terraces as mid-Wisconsinan, Sangamonian, or whatever age the evidence indicates.

Exception, however, has to be taken with Otvos' (this vol.) generalized cross-section of the Gulf Coast. Although this representation might be adequate for the Mississippi shoreline the situation in Louisiana is much more complex. Detailed study of bore holes made for the construction of causeway and bridges across Lake Pontchartrain revealed the presence of two superimposed paleosols beneath the Holocene deposits (Saucier, 1977). The upper paleosol caps a fill that contains organic materials that have produced Farmdalian $^{14}C$ dates. The base of this unit is marked by the lower paleosol which in turn tops coastal plain deposits that are "dead" to the $^{14}C$ method.

Evidently, the upper unit is the coastwise equivalent of the Waddel Bluff fill. The lower unit is probably Sangamonian and corresponds with Otvos' (this vol.) Prairie in Mississippi.

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